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ATMOSPHERIC TIDES AND ASSOCIATED MAGNETIC EFFECTS

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ATMOSPHERIC TIDES AND ASSOCIATED MAGNETIC EFFECTS

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ABSTRACT

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The dynamical theory of atmospheric tides and its relation to observations on the ground and in the upper atmosphere are reviewed briefly. The lunar tidal oscillations of atmospheric pressure observed on the ground are thought to be almost entirely due to gravitational forces whereas the solar tidal oscillations are believed to be mainly caused by solar heating. Previous theories of resonant, gravitational excitation of the solar tidal oscillations are not consistent with models of the upper atmosphere obtained from rocket measurements. Tidal oscillations at ionospheric heights are inadequately known experimentally and are inadequately explained by present theories which neglect viscuous losses and the ionized nature of the upper atmosphere and whose linear approximations break down at ionospheric height.

The dynamo theory attributes the quiet day variations of the geomagnetic field to tidal oscillations at ionospheric height. The theory has been used either to derive the magnetic variations from an assumed tidal wind system or to derive the tidal wind system from the observed magnetic variations. The results of both types of calculations lend broad support to the dynamo theory although neither of the two approaches is free of arbitrary assumptions.

The additional daily magnetic variations observed during magnetically disturbed days are particularly intense in the auroral zone. These variations are believed to be caused by some form of interaction of the solar wind with the geomagnetic field but the nature of this interaction is not yet known with certainty. Various proposed theories of these daily disturbance variationsare described and one of them is discussed in detail. This theory predicts that the generating mechanism responsible for the daily disturbance variations is present, in reduced strength, even on magnetically quiet days. Observations tend to confirm this conclusion

Author

1. Introduction

This paper will be concerned with tidal motions of the upper atmosphere and with the magnetic effects of the ionospheric currents associated with such motions. The word tide is commonly associated with the alternate rise and fall of the surface of the oceans, twice in each lunar day. These oceanic tides are believed to be caused by the gravitational forces of the moon and, to a lesser extent, of the sun. Periodic variations in atmospheric pressure, similar to the rise and fall of the ocean surface, have been observed since the 17th century, first in the tropics (where other variations are less prominent) and then at moderate latitudes. Surprisingly, the period of the observed variations of atmospheric pressure is related to the solar, and not to the lunar, day. Careful statistical studies show, however, that much weaker variations, whose period is related to the lunar day, are also present in the observed pressure data. It has become customary to refer to atmospheric pressure fluctuations, whose period is related to the solar and lunar day, as solar and lunar atmospheric tides.

The observed lunar tides in atmospheric pressure are mainly semidiurnal (they have a period of 12 lunar hours) and this is to be expected (as will be shown later) since their cause is the gravitational field of the moon. The observed solar tides in atmospheric pressure have components with periods of 24, 12, 8, 6...hours; the semidiurnal component is usually the largest. The presence of a large 24-hour component appears to suggest, however, the existence of other than gravitational excitation. The anomalously large amplitude of the solar tides in atmospheric pressure (in comparison with the lunar tides) could be interpreted as an even stronger indication of thermal rather than gravitational origin. The relative importance of gravitational and heating effects in the production of the semidiurnal component of atmospheric solar tides is a complex question that has occupied the attention of many prominent mathematical physicists since Laplace (1799, 1825). A detailed excellent review of the dynamical theory of tidal atmospheric oscillations has recently appeared (Siebert, 1961) following an earlier one by Wilkes (1949). The present account of tidal theory will therefore be relatively brief although reference will be made to the results of theoretical investigations carried out since Siebert's review.

It will become clear from the present paper that the tidal motions of the atmosphere, associated with the periodic pressure fluctuations, would not be restricted to the lower atmosphere even if the sources of their energy were so restricted. The motions of the unionized part of the atmospheric gas certainly extend to well above 100 km, while indirect effects extend to distances of several earth radii. At these great heights the atmosphere is almost fully ionized, and its motion is mainly a motion of charged particles. It will also be seen that the tidal motions produce currents in the upper atmosphere. These currents flow mainly at heights of about 90-140 km where the ions tend to move with the neutral particles as a result of collisions while the motion of the electrons (at least in directions perpendicular to the magnetic field) is almost entirely determined by the ambient electric and magnetic fields and is hardly perturbed by collisions.

The existence of upper atmospheric currents was postulated by Stewart (1882), who tried to explain the small daily variations of the geomagnetic field (the Sq variations, related to the solar day and the much weaker L variations, related to the lunar day) observed during magnetically quiet days. A similar suggestion was made considerably earlier by Gauss (1839). Stewart simply assumes that the atmosphere at great heights behaves like a moving conductor of electricity, the motion being caused by the tides. Currents are induced in the atmosphere very much like in the moving conductor of a dynamo. This explanation of the Sq and L variations is usually called the dynamo theory, and sometimes the ionospheric currents that are believed to be responsible for the magnetic S_q and L variations, are called the dynamo currents.

Stewart's theory was formulated quantitatively by Schuster (1908) and by Chapman (1919). These earlier formulations of the theory assume an isotropic conductivity in the upper atmosphere. The anisotropic nature of ionospheric conductivity is taken into account by Hirono (1952), Baker and Martyn (1952, 1953), Fejer (1953), H. Maeda (1955) and Kato (1956) among others. Their work shows that if the electron concentrations measured by ionosondes are assumed then the tidal winds that are inferred to exist at ionospheric heights from radio observations (of drifting meteor trails and of drifting diffraction patterns), are capable of producing, by dynamo action, magnetic variations of about the same order of magnitude as the observed ones.

The observed daily variations of the geomagnetic field are greatly enhanced in the polar regions, especially in the auroral zone, during magnetically disturbed days (Chapman, 1935). Attempts have been made

(Obayashi and Jacobs, 1957, H. Maeda, 1957, Swift, 1963a, 1963b) to explain the additional daily variations of the geomagnetic field (the Ds variations) that occur during disturbed days, by modified tidal dynamo theories in which enhanced E-region ionization densities (and therefore conductivities) are assumed to exist in the auroral zone. The results of such attempts show that the observed Ds variations cannot be explained by a dynamo theory without rather artificial assumptions about the atmospheric wind system that drives the currents.

In other, more recent theories of the Ds variations (Chamberlain, 1961; Fejer, 1961, 1963a; and Kern, 1961) the energetic charged particles of the radiation belts, which are trapped in a geomagnetic field distorted by the solar wind, play an essential part. It is natural to include these theories in a discussion of the magnetic effects caused by atmospheric tides. It is true that some of the mechanisms proposed to explain the Ds variations are unrelated to the tidal atmospheric pressure fluctuations observed on the ground, but it is not easy to separate the ionospheric currents caused by tidal winds (associated with the pressure fluctuations on the ground), without a theoretical understanding of both effects. Moreover, if the meaning of the adjective "tidal" is extended to describe motions of the magnetosphere, and if a particle stream from the sun is added to electromagnetic radiation and gravity, as yet another possible cause of solar tides, then any mechanism suggested for the explanation of the Ds variations is necessarily a solar tidal mechanism. Although the word "tidal" will not be used in this wider sense in the present article, all periodic magnetic variations whose period is related to the solar, or lunar, day will be discussed and the theories proposed for their explanation will be considered.

2. Gravitational Equilibrium Tides and the Observed Pressure Fluctuations

When the earth moves under the influence of another astronomical body such as the sun or the moon, its center of mass moves with an acceleration $\underline{a}$ that can be calculated by assuming that all the forces acting on the body act at the center of mass and that the whole mass is concentrated there. In addition, the earth rotates about its center of mass with an angular velocity $\underline{\omega}$ which may be regarded as constant for present purposes.

If terrestrial motions are studied from a coordinate system fixed with respect to the earth and with its origin at the center of mass, then three "inertial forces" have to be added to any other forces acting on a hypothetical point mass m. These are the "inertial" force $- m\underline{a}$, the centrifugal force $m\underline{\omega}\times(\underline{r}\times\underline{\omega})$ and the Coriolis force $2m\underline{v}\times\underline{\omega}$ where $\underline{v}$ is the velocity of the point mass and $\underline{r}$ is the radius vector.

It is of interest to examine first the equilibrium distribution of the oceans and the atmosphere under the influence of the forces acting on them; in such a calculation the earth's angular velocity is assumed to be infinitely small and the centrifugal and Coriolis forces are neglected. The three remaining external forces acting on a point mass m are then the gravitational attractions of the earth and of the perturbing astronomical body and the inertial force $-m\underline{a}$. One of these forces, the gravitational attraction of the earth remains fixed with respect to the earth, and thus does not cause any changes in the equilibrium distribution of the oceans and the atmosphere. The inertial force $-m\underline{a}$ and the gravitational attraction of the perturbing body remain approximately fixed with respect to the perturbing body, and thus appear to rotate when viewed from a system fixed with respect to the earth. These two forces are very nearly equal in magnitude and opposite in direction for a hypothetical point mass situated at the center of the earth. The resultant of the two forces does not, however, quite vanish on the surface of the earth, as illustrated schematically by Fig. 1 for points in the equatorial plane; this resultant force is responsible for the gravitational tidal motions. Figure 1 shows that, if the earth's axis of rotation is taken to be perpendicular to the earth-sun line, then the tidal force goes through two cycles of rotation while the earth goes round once. The variation of the tidal force is thus semidiurnal in nature; in reality, a smaller diurnal component is also present because the earth's axis of rotation is not quite perpendicular to the sun-earth direction.

Since both the attractive force of the perturbing body and the uniform force field $-m\underline{a}$ may be expressed as gradients of a potential, it is more convenient to calculate the tidal potential rather than the tidal force. A detailed harmonic analysis of the calculated tidal potential was carried out by Doodson (1922). If the tides were caused by gravitation alone, and if the atmosphere would assume an equilibrium distribution corresponding to the tidal potential at all times, then at the equator the amplitude of the average lunar semidiurnal variation of atmospheric pressure would be about 0.022 mm of mercury; the solar semidiurnal variation would have an amplitude of 0.010 mm of mercury. (Smaller diurnal variations would also be expected on account of the inclination of the earth's eqatorial plane with respect to the ecliptic.) The observed average amplitudes at the equator are about 0.9 mm of mercury for the solar, and 0.055 mm for the lunar, semidiurnal pressure fluctuations. The detailed geographical distribution of the calculated equilibrium tidal pressure fluctuations and of the observed fluctuations is given by Siebert (1961), and the reader is referred to his review article for further detail.

It is clear that the observed lunar semidiurnal pressure fluctuations exceed the gravitational equilibrium tide by a factor of about 2.5, while for the solar semidiurnal pressure fluctuation the value of the same factor is about 90.

3. Gravitational and Thermal Excitation of Atmospheric Oscillations

3.1 Basic Equations

Laplace (1799, 1825) realized that the simple equilibrium theory of atmospheric tides cannot explain the tidal pressure fluctuations observed on the ground and therefore formulated his dynamical theory of the tides. His work was further developed by Margules (1892, 1893), Hough (1897, 1898), Lamb (1910), Taylor (1936) and Pekeris (1937, 1939). More recent work by Weekes and Wilkes (1947), Jacchia and Kopal (1952), Siebert (1954) and Small and Butler (1961, 1963) applies the theory to different and increasingly realistic atmospheric models in attempts to bring the theory into better agreement with observations.

Certain approximations are common in all the above theories. The earth is assumed to be a perfect sphere of radius R that rotates with a constant angular velocity $\underline{\omega}$. The change of magnitude of the gravitational acceleration $\underline{g}$ with height and geographical position is neglected. The static pressure p_o, the density ρ_o, the temperature T_o and the scale height $H = KT_o/mg$ (where K is Boltzmann's constant) of the atmosphere are assumed to depend only on the height z and not on the colatitude θ and the longitude ϕ. The mean molecular mass m is taken to be independent of height; this is probably true below the lower boundary of the ionosphere. Non-linear terms are neglected in the equations of motion; this is certainly not permissible at heights above about 100 km, where the tidal pressure fluctuations become comparable to the static pressure p_o. Vertical accelerations are neglected in the equations of motion since the motions are largely horizontal. Dissipative energy losses are neglected. A more detailed discussion of these approximations and their range of validity may be found in Siebert's (1961) review.

With the above approximations, the equations of motion of fluid mechanics lead to a partial differential equation which can be solved by separation of variables. The details of the analysis are given by Siebert (1961). The solutions are obtained in terms of orthogonal eigenfunctions $\psi_{1,m}$, corresponding to eigen values $h_{1,m}$ of the differential equation

$$F\Psi + (4R^2\omega^2/gh)\Psi = 0 \qquad (1)$$

Here F denotes the differential operator

$$F = \frac{1}{\sin\theta}\frac{\partial}{\partial\theta}\left(\frac{\sin\theta}{f^2-\cos^2\theta}\frac{\partial}{\partial\theta}\right) + \frac{1}{f^2-\cos^2\theta}\left[\frac{i}{f}\frac{f^2+\cos^2\theta}{f^2-\cos^2\theta}\frac{\partial}{\partial\phi} + \frac{1}{\sin^2\theta}\frac{\partial^2}{\partial\phi^2}\right] \qquad (2)$$

where $f = \sigma/2\omega$ and $2\pi/\sigma$ is the period of the tidal oscillation. The unknown divergence $\chi = \text{div}\,\underline{v}$ of the velocity $\underline{v}$ of the atmosphere may then be written in the form of an expansion in terms of the orthogonal eigenfunctions $\psi_{1,m}$ in the form

$$\chi = \sum_{l,m} \chi_{l,m}(z)\,\Psi_{l,m}(\theta,\phi)\,e^{i\sigma t} \qquad (3)$$

where $\chi_{1,m}(z)$ are complex functions of the height to be determined. The tidal potential $\Omega(z,\theta,\phi)$ and the amount of heat $J(z,\theta,\phi)$ absorbed by unit volume of the atmosphere in unit time may be similarly expanded. The unknown functions $\chi_{1,m}(z)$ may then be shown to be determined from the differential equations

$$\frac{d^2 y_{l,m}}{dx^2} - \frac{1}{4}\left\{1 - \frac{4}{h_{l,m}}\left[\kappa H(x) + \frac{dH}{dx}\right]\right\} y_{l,m}(x) = \frac{\kappa J_{l,m}(x)}{\gamma g h_{l,m}}\,e^{-x/2} \qquad (4)$$

where $H = kT/mg$ is the scale height, k is Boltzmann's constant, T is the temperature, $\lambda = c_p/c_v$ is the specific heat ratio, and where

$$\kappa = (\gamma - 1)/\gamma \tag{5}$$

$$x = \int_0^z dz/H(z) \tag{6}$$

$$y_{\ell,m}(x)\, e^{x/2} = \chi_{\ell,m}(z) - \frac{\kappa J_{\ell,m}(z)}{g H(z)} \tag{7}$$

It should be noted that equation (4) closely resembles the wave equation. If the temperature is low (H is small) and the temperature gradient is negative, then the expression in the curly bracket can become negative and the waves are then evanescent. At the ground the vertical component of the velocity must vanish; the resulting boundary condition assumes the form

$$\left[\frac{dy_{\ell,m}}{dx} + \left(\frac{H}{h_{\ell,m}} - \frac{1}{2}\right) y_{\ell,m}\right]_{x=0} = \frac{i\sigma}{\gamma g h_{\ell,m}}\, \Omega_{\ell,m}(0) \tag{8}$$

Another boundary condition has to be satisfied at infinity (at great heights). If dissipative losses were taken into account by the theory, then this boundary condition could be formulated in terms of the vanishing of the oscillation amplitude at infinity. Since such losses are not included in the theory, the atmosphere at great heights is usually chosen so that it can support propagating waves (described by equation 4) and the boundary condition is then taken to be that waves at great height must be outgoing (i.e., the amplitude of incoming waves must vanish).

Some confusion exists in the literature about the boundary condition at infinity when the chosen atmospheric model can support propagating waves at great heights. This boundary condition is often stated in the form that the outflowing energy flux and therefore the imaginary part of y^* (dy/dx) must be positive (the asterisk indicates the conjugate complex value). Such a statement is not sufficient for the formulation of a boundary condition at infinity; in effect it merely insists that the energy flux of the outgoing wave should be greater (without specifying how much greater) than the energy flux of the incoming wave. The correct boundary condition must insist on the vanishing of the incoming wave; the calculations of Weekes and Wilkes (1947), Jacchia and Kopal (1952) and Small and Butler (1963) all assume, in effect, that the amplitude of the incoming wave vanishes.

Equations (1-8) formally describe the method by which solutions can be obtained for the tidal oscillations of a given model atmosphere, excited by given tidal forces, usually by numerical methods. The other quantities of interest, such as the pressure and the conponents of the velocity, may then be expressed in terms of the solution for χ; the relevant equations are given by Siebert (1961).

3.2 Internal Gravity Waves

The dynamical problem of tidal oscillations is essentially one of wave propagation in a spherically stratified medium; the waves are excited by gravitational and thermal forces. The frequency of the waves is very low; their period is of the order of hours.

The nature of propagating waves in an idealized plane-stratified atmosphere whose temperature is independent of the height, changes considerably as the period of the waves is gradually increased. Conventional sound waves cease to propagate when their wave length becomes comparable to or larger than the scale height of the atmosphere. More precisely, waves whose period falls within a critical time interval (which is between about 4.4 and 4.9 minutes for a scale height of 6 km and a gravitational acceleration of 9.5 m/sec^2), cannot propagate at all; waves with shorter periods resemble conventional sound waves, whereas waves with longer periods are essentially different from sound waves and are usually called internal gravity waves. The following brief discussion of the propagation properties of these waves is based on a recent paper by Hines (1960).

In contrast to high frequency sound waves whose wave length is determined by the frequency alone, the wave length of gravity waves, with periods substantially in excess of the critical 4.9 minutes, strongly depends on their direction of phase propagation. It can have any value smaller than a certain critical value which is close to the wave length that a sound wave of the same frequency would have. For all wave lengths that are substantially smaller than this critical value, the direction of phase propagation is very nearly vertical and the direction of the group velocity (or energy propagation) is very nearly horizontal; the small vertical component of the group velocity points in a direction opposite to that of the vertical component of the phase velocity.

Another important difference between internal gravity waves of long period and sound waves of short period lies in the direction of oscillations of the medium. In sound waves the direction of oscillation is nearly parallel to the direction of phase and group propagation, whereas in internal gravity waves, the direction of oscillation is approximately horizontal and therefore is nearly perpendicular to the direction of phase propagation and at the same time nearly parallel to the direction of group propagation.

A further interesting property of internal gravity waves is the constancy of the product $\rho_0 \overline{u^2}$ where ρ_0 is the density of the medium and $\overline{u^2}$ is the mean square value of the velocity of oscillation. The velocity of the oscillations can therefore be many orders of magnitude greater at ionospheric heights than on the ground.

3.3 Free Oscillations of the Atmosphere

In effect equations (1-8) mathematically describe the excitation of long period internal gravity waves in the earth's non-uniform atmosphere by gravitational and thermal tidal forces. Before the calculation of these forced oscillations of the atmosphere, it is of interest to inquire whether free oscillations of the atmosphere are possible. The condition for such free oscillations is that an eigenvalue $h_{1,m}$ of equation (1) should be equal to an eigenvalue h_n of equation (4), after setting $\Omega = J = 0$ and omitting the subscripts 1,m. The eigenvalues h_n of equation (4) may be given a physical interpretation; it may be shown (Taylor 1936) that $(gh_m)^{\frac{1}{2}}$ is the velocity of propagation of long internal gravity waves spreading out horizontally from explosions such as the Krakatoa eruption in 1883 or the recent high altitude thermonuclear explosions. Since $(gh_n)^{\frac{1}{2}}$ is also the velocity of long waves in an ocean of depth h_n, the eigenvalues h_n could, in a sense, be regarded as the equivalent depths of the atmosphere. It appears puzzling at first that the atmosphere could have more than one equivalent depth, corresponding to more than one horizontal velocity of propagation. A situation rather similar to that in a wave guide which can support more than one propagating mode may, however, occur in the atmosphere.

If one interprets equation (4) as a wave equation, then wave propagation is possible only when the coefficient of $y_{1,m}$ in the curly bracket is positive; otherwise the waves are evanescent. In the earth's atmosphere, for a given value of h, this coefficient could be negative near the tropopause and near the mesopause (where there are barriers) and positive near the mesopeak (where there may be a duct) and in the thermosphere. It may be shown that two eigenvalues h_n may exist for certain atmospheric models such as the one proposed by Weekes and Wilkes (1947) and shown by Fig. 2; one of these corresponds to propagation trapped below the tropopause while the other corresponds to propagation trapped below the mesopause. While the first of these eigenvalues does not depend greatly on the choice of an atmospheric model (since the atmosphere is rather well known and simply behaved below the tropopause), the second eigenvalue strongly depends on the temperature profile assumed for the mesosphere and may even be altogether absent if the temperature maximum at the mesopeak is not sufficiently high. (On the other hand atmospheric models with more than two eigenvalues have also been tentatively suggested by Weekes and Wilkes.)

The eigenvalues $h_{l,m}$ may also be given a physical interpretation. The free oscillations of a uniform ocean are governed by a differential equation for the perturbation in the surface elevation that only differs from equation (1) in that h is replaced by the depth of the ocean. The eigenvalues of the equation, for a given period of oscillation, represent those (resonant) depths of the ocean for which free oscillation of that period are possible. There is a double infinity of such eigenvalues $h_{l,m}$ (corresponding to the number of latitudinal and longitudinal nodes l and m of the oscillations) each of which is only a function of the period. These eigenvalues, which have been computed by Hough (1897, 1898), not only represent the resonant depths of a uniform ocean but also the resonant "equivalent depths" of a horizontally uniform atmosphere. For the (2,2) mode (l = 2, m = 2) which resembles the observed semidiurnal pressure fluctuations of the earth's atmosphere and for a period of half a solar day, the resonant depth is 7.9 km; for half a lunar day the resonant depth is 7.1 km. It should be stressed that these values apply equally to an ocean of uniform depth and to the atmosphere and do not depend on the vertical atmospheric structure as long as it does not vary with geographical position.

3.4 The Resonance Theory of Tidal Oscillations

Resonant excitation of the atmosphere by the semidiurnal solar tidal forces could take place if one of the eigenvalues h_n of equation (4) (one of the equivalent depths of the atmosphere) were close to 7.9 km; such a resonance could explain the relatively large factor by which the observed solar semidiurnal pressure fluctuations exceed both the gravitational equilibrium tide and the observed lunar semidiurnal pressure fluctuations without invoking thermal excitation. Lamb (1910) pointed out the possibility of such a resonant excitation of the solar tides in the (2,2) mode. He also pointed out that the resonant period would have to be within a few minutes of 12 solar hours to obtain a sufficiently great amplification of the solar tides over equilibrium tides and over the lunar tides, assuming gravitational excitation alone. It seems difficult to believe that such accurate tuning is permanently maintained. Taylor (1936) pointed out another difficulty in connection with the resonance theory; the equivalent depth of the atmosphere could be calculated from the measured velocity of propagation of very long internal gravity waves from the Krakatoa explosion, since this velocity must be equal to $(gh_n)^{\frac{1}{2}}$, as mentioned in the previous section. An equivalent depth of just over 10 km is obtained in this manner, which is very much greater than the required 7.9 km.

Taylor indicated, however, a possible way out of this difficulty; the atmosphere may have more than one equivalent depth and only one of these is obtained from the first arrival of waves from explosions. This idea was investigated in greater detail by Pekeris (1937, 1939) and by Weekes and Wilkes (1947). Figure 2 shows the temperature distribution assumed by Weekes and Wilkes; the 1962 U. S. Standard Atmosphere is shown for comparison. Weekes and Wilkes (1947) showed that the resonant depths of 10.4 km and 7.9 km are obtained for their model atmosphere. The first value could explain the observations of wave propagation from the Krakatoa explosion; the second value could account for the resonant amplification of the solar tide. Weekes and Wilkes have shown that the exact value of this second equivalent height and the sharpness of the resonance associated with it depend very sensitively on the assumed atmospheric model. (The damping of the resonance is introduced by the assumption of an insufficiently thick and therefore imperfect mesospheric barrier and an isothermal top above it, which allows the waves to escape to infinity.)

As mentioned before, it seems unlikely that such an accurate tuning of the atmosphere could be maintained at all times. Moreover, Jacchia and Kopal (1952) have shown that the second equivalent height of the atmosphere disappears altogether if a more realistic atmospheric model based on the results of rocket measurements (essentially similar to the 1962 U. S. Standard Atmosphere) is assumed and that resonant excitation by predominantly gravitational forces cannot be invoked to explain the observed pressure fluctuations.

3.5 Thermal Excitation of Tidal Oscillations

Thermal excitation of the solar semidiurnal tidal oscillations has been considered by Siebert (1957, 1961) and Small and Butler (1961, 1963). The thermal processes usually considered are turbulent mass exchange or eddy conductivity in the lower atmosphere and heating processes arising out of absorption by atmospheric gases of the radiation from the sun and the earth. The absorption of solar radiation by atmospheric gases could provide a large semidiurnal component of the tidal excitation, although not quite as large as the diurnal component.

Siebert (1957) shows that the absorption of solar radiation by water vapor is 10 times more effective than eddy conductivity in exciting semi-diurnal pressure oscillations and could account for a substantial fraction of the observed amplitudes. Small and Butler (1961), using a more realistic temperature profile than Siebert, calculated the contribution to the solar semidiurnal oscillation at ground level arising out of the direct absorption of insolation by ozone and found that it was greater than all other previously considered contributions. Small and Butler claim that the result obtained by them for excitation by the absorption of insolation by ozone together with Siebert's result for absorption by water vapor, can adequately account for 80 per cent of the solar semidiurnal pressure amplitude observed on the ground. Small and Butler further point out that the absorption of insolation by ozone cannot account for the observed solar diurnal pressure fluctuations.

The reason for this is the small resonant depth of 0.63 km (the eigenvalue h_{11} of equation 1 appropriate for the diurnal oscillations) and the resulting small vertical "wave length" (determined by equation 4) of the oscillations; since the excitation by the absorption of insolation by ozone occurs over a large height range, the excited waves do not add in phase but rather tend to nearly cancel. Presumably the observed diurnal pressure fluctuations are excited by some other form of heating that occurs over a narrower height range.

The variation of the amplitude and phase of the semidiurnal pressure fluctuations with height obtained from the calculation of Weekes and Wilkes (1947) and of Small and Butler (1963) are shown by Figures 3a and 3b. Comparison with Figure 2 shows that the oscillation amplitude grows rapidly with increasing height in the regions where the temperature is high and the temperature gradient is not negative (where"wave propagation"is possible) but that the amplitude grows very slowly with height in those regions where the temperature is low and where the temperature gradient is negative; this type of behavior would be expected from equation (4). The similarity of the curves resulting from the two theories is remarkable in view of the radically different nature of the assumptions: a resonant atmosphere excited by gravitational tidal forces on the one hand and a non-resonant atmosphere excited by the absorption of solar radiation on the other hand. The rather large increase in relative pressure amplitude p/p_o and therefore in velocity with height results in tidal oscillation velocities which are about two orders of magnitude greater at ionospheric heights (at, say, 100 km) than on the ground. Figure 3a also shows that at heights much greater than 100 km linear theory must break down since the predicted pressure perturbations become comparable with the static pressure.

It should be stressed that the increase in the velocity of oscillation with height is not a resonant effect but a natural property of internal gravity waves. The increase would be much greater if it were not for the regions of evanescence. The density at 100 km is less than a millionth of the density at sea level, and therefore the velocity of oscillation in an isothermal atmosphere would be expected to be more than a 1000 times greater at a height of 100 km than at sea level. It should also be noted that a null and a phase reversal of the semidiurnal pressure fluctuations at a height of about 30 km is predicted by both theories.

3.6 Comparison with Observations

The predictions of the theory for semidiurnal wind velocities and their variation with height in the 80 - 100 km height region of the ionosphere, have been compared with the measured velocities of drifting meteor trails (Greenhow and Neufeld 1956). Reasonable agreement is obtained, if yearly averages of the experimental data are considered. There is, however, a very large seasonal variation in the observed data which is not explained by the theory. A more detailed discussion of the drifting meteor trail observations and their relation to the theory may be found in an excellent review paper on upper atmospheric motions by Hines (1963a), who points out that semidiurnal modes with smaller resonant depths than the dominant (2,2) mode, whose amplitude is relatively weak but measurable at sea level, may be less attenuated in the evanescent regions of the atmosphere and may therefore play a dominant part at ionospheric levels. The predominance of the diurnal over the semi-diurnal variations in some of the drifting meteor trail observations may similarly be due to a smaller attenuation of the diurnal mode whose resonant depth is less than 1 km.

Less direct indications of upper atmospheric winds at ionospheric heights are obtained from the observation of drifting diffraction patterns, associated with the fading of radio waves, with three closely spaced receivers.

(Briggs and Spencer 1954, Briggs 1960). The interpretation of such observations is difficult because the irregularities in ionization generally do not move with the mean velocity of the neutral particles (Clemmow, Johnson and Weekes 1955).

An even less direct indication of upper atmospheric winds is the presence of lunar semidiurnal variations in the ionospheric layers (Martyn 1950), revealed by ionosonde data.

Further discussion of these indirect indications of motions (with more references) will be found in the review by Hines (1963a).

3.7 Discussion

At the present stage of its development, the dynamical theory of tidal atmospheric oscillations appears to provide a reasonably good explanation of the semidiurnal pressure fluctuations observed on the gound. The lunar fluctuations are of purely gravitational origin; Small and Butler's calculations result in an amplification of 2.5 over the equilibrium tides, in good agreement with the observations. The solar semidiurnal fluctuations observed on the gound are apparently due to excitation by the absorption of solar radiation by water vapor and by ozone, and only a very small fraction of them is due to gravitational forces. The solar diurnal pressure fluctuations observed on the ground are less well explained and there are even more gaps both in our knowledge of tidal atmospheric oscillations at ionospheric heights and in their theoretical explanation. This is unfortunate since only those oscillations of the atmosphere that occur at ionospheric heights can produce the magnetic effects which are the subject of the present review.

4. The Dynamo Theory

4.1 The Conductivities

When the oscillations of the upper atmosphere extend to ionospheric heights, they involve the motion of charged as well as of neutral particles.

In the presence of the geomagnetic field, the mean velocities of the neutral, the positively charged and the negatively charged particles are generally not the same and ionospheric currents result. In a qualitative manner, these currents may be said to be induced by the motion of conducting air across the geomagnetic field just as the current in the winding of a dynamo is induced by the motion of a conductor in a magnetic field. This analogy cannot, however, be pursued too far because in the presence of a magnetic field, the conductivity of a partially ionized gas is not isotropic. The relations derived by Cowling (1932) between the total electric field $\underset{\sim}{E}$ and the current density $\underset{\sim}{J}$ are most easily expressed in terms of components $\underset{\sim}{E}_{\parallel}$ and $\underset{\sim}{J}_{\parallel}$ parallel to the geomagnetic induction $\underset{\sim}{B}_0$ and vector components $\underset{\sim}{E}_{\perp}$ and $\underset{\sim}{J}_{\perp}$ normal to $\underset{\sim}{B}_0$.

$$\underset{\sim}{J}_{\parallel} = \sigma_0 \underset{\sim}{E}_{\parallel} \quad , \qquad \underset{\sim}{J}_{\perp} = \sigma_1 \underset{\sim}{E}_{\perp} + \sigma_2 \underset{\sim}{B}_0 \times \underset{\sim}{E}_{\perp} / B_0 \tag{9}$$

The various conductivities σ_0 ('longitudinal'), σ_1 ('Pedersen') and σ_2 ('Hall') are given by the expressions (derived in Chapman and Bartels, 1940, pp. 531-535).

$$\sigma_0 = \frac{Ne}{B_0}\left(\frac{\omega_i}{\nu_i} + \frac{\omega_e}{\nu_e}\right) \tag{10}$$

$$\sigma_1 = \frac{Ne}{B_0}\left(\frac{\omega_i \nu_i}{\omega_i^2 + \nu_i^2} + \frac{\omega_e \nu_e}{\omega_e^2 + \nu_e^2}\right) \tag{11}$$

$$\sigma_2 = \frac{Ne}{B_0}\left(\frac{\omega_e^2}{\omega_e^2 + \nu_e^2} - \frac{\omega_i^2}{\omega_i^2 + \nu_i^2}\right) \tag{12}$$

where N is the number density of free electrons, e is the absolute value of their charge, $\omega_i = eB_0/m_i$, $\omega_e = eB_0/m_e$ are the gyrofrequencies (m_i, m_e are the masses) and ν_i, ν_e are the collision frequencies (with neutral particles) of ions and electrons. Figure 4 shows the quantities $\sigma_0 B_0/Ne$, $\sigma_1 B_0/Ne$ and $\sigma_2 B_0/Ne$, which are proportional to the conductivities per unit electron density, as functions of the height, for representative values of the collision frequencies and gyrofrequencies at a moderate latitude.

Figure 4 shows that in the F region, well above 160 km, σ_0' is extremely large whereas σ_1' and σ_2' are very small; the inequality $\sigma_0' \gg \sigma_1' \gg \sigma_2'$ is satisfied. The Hall conductivity σ_2' is much larger than the Pedersen conductivity σ_1' in the 90 - 130 km region although σ_0' is still larger than either σ_1' or σ_2'. At very low heights, below 70 km, $\sigma_0' \sim \sigma_1'$ and they are both much larger than σ_2'; the magnetic field here has little effect and the ionospheric conductivity is isotropic and very small. A more detailed discussion of ionospheric conductivity will be found in a review by Chapman (1956).

The total electric field E consists of two distinct parts. One of these is due to the motion of the atmosphere with a velocity $\underset{\sim}{u}$; equations (9) must then be applied in a system moving with the atmosphere and in this system an (induced) electric field $E_i = \underset{\sim}{u} \times \underset{\sim}{B}_0$ must be added to the "polarization" electric field $\underset{\sim}{E}_p$ that is seen from a stationary system and that results from an accumulation of polarization charges. The electric field $\underset{\sim}{E}_i$ is often called the dynamo electric field.

In the F region and above the current is strongly inhibited from flowing across a field lines by the low values of σ_1' and σ_2' and therefore the steady state current carried along a tube of force must be independent of position. If symmetry about the geomagnetic equator is assumed, both of the wind system and of the conductivity, then no currents flow along the tubes of force in the F region and above. The currents then flow in a relatively thin spherical shell and the vertical current density therefore vanishes approximately at all heights. The equation that expresses the vanishing of the vertical component of the current density can be used to eliminate the vertical component of the total electric field from the components of the vector equation (9), written in a coordinate system whose x, y, z axes point to the south, the east and the zenith respectively.

The equations:

$$J_x = \sigma_{xx} E_x + \sigma_{xy} E_y \tag{13}$$

$$J_y = -\sigma_{xy} E_x + \sigma_{yy} E_y \tag{14}$$

are obtained in this manner, where the conductivities $\sigma_{xx}, \sigma_{xy}, \sigma_{yy}$ are functions of σ_0, σ_1, σ_2, and the angle of inclination I; they are given by the equations:

$$\sigma_{xx} = K^{-1} \sigma_1 \sigma_0 \tag{15}$$

$$\sigma_{xy} = K^{-1} \sigma_2 \sigma_0 \sin I \tag{16}$$

$$\sigma_{yy} = K^{-1} (\sigma_1 \sigma_0 \sin^2 I + \sigma_1 \sigma_3 \cos^2 I) \tag{17}$$

$$K = \sigma_1 \cos^2 I + \sigma_0 \sin^2 I \tag{18}$$

$$\sigma_3 = \sigma_1 + \sigma_2^{\,2} / \sigma_1 \tag{19}$$

The height integrated conductivities $\int \sigma_{xx} dz, \int \sigma_{xy} dz$ and $\int \sigma_{yy} dz$ play an important part in the dynamo theory and are shown by Figure 5 as functions of the latitude, for typical midday conditions. Figure 5 shows that $\int \sigma_{xy} dz$ predominates over most parts of the earth but that at the magnetic dip equator $\int \sigma_{xy} dz$ vanishes and the height integrated conductivities $\int \sigma_{xx} dz$, $\int \sigma_{yy} dz$ assume very large values. These large values are believed to be the cause of the observed very high values of the magnetic S_q and L variations near the magnetic dip equator.

4.2 Differential Equations of the Dynamo Theory

It is convenient to write equations (13) and (14) in the two dimensional vector form

$$\underset{\sim}{J}_h = \underset{\approx}{\sigma}_h \cdot \left[E_{ph} + (\underset{\sim}{u} \times B_0)_h \right] \tag{20}$$

where $\underset{\sim}{J}_h$, $\underset{\sim}{E}_h$, $\underset{\sim}{E}_{ph}$, $(\underset{\sim}{u} \times \underset{\sim}{B}_o)_h$ are the horizontal components of the current density $\underset{\sim}{J}$, the total electric field $\underset{\sim}{E}$, the polarization electric field $\underset{\sim}{E}_p$ and the induced electric field $\underset{\sim}{E}_i = \underset{\sim}{u} \times \underset{\sim}{B}_o$ respectively. Over the height region 90-160 km, where most of the currents flow, the potential differences due to vertical electric fields are small compared to the potential differences that exist in horizontal directions over distances of thousands of kilometers. It is therefore reasonable to assume that $-E_{ph}$ is independent of the height and may be expressed as the gradient of a two dimensional potential function Ψ on the current carrying ionospheric shell.

In the quasi-stationary state the (two dimensional) divergence of the height integrated current density div ($\int \underset{\sim}{J}_h \, dz$) must vanish if the very small currents, required to build up the polarization charges, are neglected. With the aid of equation (20) this condition leads to:

$$\mathrm{div}\left[\left(\int \underset{\approx}{\sigma}_h \, dz\right) . \, \mathrm{grad}\, \Psi\right] = \mathrm{div}\left[\int \underset{\approx}{\sigma}_h . (\underset{\sim}{u} \times B_o)_h \, dz\right] \tag{21}$$

Equation (21) is a partial differential equation for Ψ, as a function of latitude and local time. If the conductivities and wind velocities are known, then equation (21) may be solved for Ψ, by numerical methods. The differential equation has been further simplified by some authors (Fejer, 1953, Lucas, 1954) who neglected the diurnal variation of electron density and assumed that $\underset{\approx}{\sigma}_h$ is a function of the latitude and the height only. After a separation of variables equation (21) then becomes an ordinary differential equation, with the latitude as the independent variable. The further assumption was made by the above authors that the wind velocity is independent of height and is proportional to the wind velocity system derived from the dominant (2,2) mode of semidiurnal pressure fluctuations observed on the ground. These pressure fluctuations and the velocities derived from them by Bartels (1928) are shown by Figure 6, The factor of proportionality was regarded as an unknown constant whose value was so adjusted that the strength of the computed current system became comparable to the semidiurnal part of the daytime current system derived from the observed S_q magnetic variations and shown by Figure 7a. A factor of 60 is obtained in this manner. The ratio between the semidiurnal wind velocities observed in the ionosphere (Greenhow and Neufeld 1955, 1956, Briggs and Spencer 1954, Briggs 1960, Hines 1963a)and the velocities of Figure 6 is not very different from this factor.

H. Maeda (1955) used a rather different approach in which the diurnal variation of the conductivity was taken into account. He regarded the current system derived from the observed Sq variations and the conductivity (known from ionosonde observations) as given and the wind velocities as unknown. He assumed that the wind velocities are independent of the height and may be represented by a velocity potential and solved equation (20) for the unknown velocity potential. Kato (1956) has carried out a similar analysis in which the Coriolis forces, neglected by Maeda, were taken into account and arrived at similar results. The diurnal and semidiurnal components of the wind system deduced by Kato are shown by Figure 8. The large diurnal, as compared to semidiurnal component of this wind system is of particular interest and shows that the large diurnal component of the observed Sq magnetic variations cannot be explained by the diurnal variation of conductivity a lone. Kato's work suggests that the winds responsible for the dynamo currents are prominently diurnal and not semidiurnal.

This suggestion appears to be supported by other evidence. Although the semidiurnal component of the wind velocities derived from meteor observations by radar at Jodrell Bank, England (Greenhow and Neufeld, 1956) is much larger than the diurnal component, it is significant (Briggs, 1960) that there is no large seasonal variation in the amplitude and phase of the observed Sq magnetic variations inspite of the presence of such variations in the semidiurnal component of the observed wind velocity. It is also significant that the diurnal component of the wind velocity was found to be larger than the semidiurnal component in Australian meteor observations (Elford and Robertson, 1953). Nevertheless some reservations must be made about the validity of Maeda's and Kato's methods of derivation of their atmospheric wind systems. First of all, the assumption of a wind velocity independent of the height is unrealistic. Yet another arbitrary tacit assumption is made by eliminating the steady component from the magnetic variation data. A steady component may well be present in the actual ionospheric current system, generated by periodic air motions, on account of the diurnal variation of conductivity. As Risbeth points out (private communication) this arbitrary assumption may lead to the introduction of a spurious diurnal component into the calculated wind system.

4.3 Discussion. Motions in the Magnetosphere

In spite of its successful rough explanation of the observed Sq magnetic variations in terms of tidal oscillations of the atmosphere no accurate and realistic formulation of the dynamo theory exists at present. This is not really surprising since our experimental and theoretical knowledge of tidal oscillations at ionospheric height is still insufficient. Further experimental and theoretical work will have to fill in the gaps in our present knowledge. It should be stressed here that an independent theoretical treatment of the tidal oscillations and of the dynamo theory neglects the reaction of the motions of the charged particles of the atmosphere on the motion of the neutral particles. Hines (1963a) points out that this reaction may be very significant at levels above 100 km and it is to be hoped that future theoretical work will take it into consideration. Another effect, that has been neglected in previous formulations of the dynamo theory, is the distortion of the Sq current system by the interaction of the polarization electric field with energetic trapped protons. This distortion, which is expected to be greatest in the auroral region, is discussed in Section 5.

An indirect consequence of the dynamo theory should be stressed here. It has already been pointed out that the magnetic lines of force are very nearly lines of equal potential. The polarization potential ψ on the ionospheric shell is therefore carried into the magnetosphere by the lines of force. The resulting electric field $\underset{\sim}{E}$ is perpendicular to the magnetic induction $\underset{\sim}{B}_0$ at all points of the magnetosphere and leads to a drift of the magnetospheric plasma. The velocity component perpendicular to the magnetic field of this drift is $\underset{\sim}{E} \times \underset{\sim}{B}_0 / B_0^2$. In these magnetospheric plasma motions any ionized matter which at one time occupies a tube of force, continues at all later times to occupy a tube of force of equal flux content. The earth's magnetic field remains almost unaltered by the motions. The tidal motions excited from below therefore not only cause currents to flow in the ionosphere, mainly in the E region, but also generate plasma motions in the magnetosphere, including the F region (Martyn 1947).

5. Theories of the Daily Disturbance Variations

5.1 The D_{st} and the D_s Variations

When the solar daily magnetic variations Sq, averaged for the five International Magnetically Quiet Days of the month are subtracted from the deviations of the field observed at other times, the resulting residual magnetic variations, D, are called the disturbance variations. These variations have a periodic component Ds whose period is one solar day. Many different notations are used for these disturbance daily variations, depending on the statistical methods used for their derivation; for simplicity only the notation Ds is used here. It is usual to represent the daily disturbance variations Ds by a current system that flows in the E region just as the Sq variations were represented by the current system in Figure 7. (Such a representation does not imply that the currents actually flow in the E region. Although currents in the E region are believed to play a predominant part in the production of both the Sq and the Ds variations, magnetospheric currents are also thought to be involved in both cases. In the case of the Ds variations the presence of magnetospheric currents is believed to be essential.) A more detailed description of the different geomagnetic variations will be found in Geomagnetism (Chapman and Bartels, 1940) and in reviews by Vestine (1960) and by Chapman (1962).

Chapman (1935) deduced the Ds variation from the observed disturbance variation D for 40 magnetic storms of moderate intensity. He identified the time of commencement of each storm (at least to the nearest hour) and then determined the hourly mean value of the disturbance D as a function of latitude, local time, and storm time (defined as the time measured from commencement). He then calculated the average value Dst of D over the 24 hours of local time, for a given storm time, and called it the storm-time variation. The daily disturbance variation Ds for a given storm time was then obtained as a function of local time after the subtraction of Dst from the mean value of D, for each value of the storm time. It is then found that Dst usually predominates at low latitudes and Ds predominates at high latitudes, especially in the auroral zone.

Figure 9 shows the Ds current system derived in this manner by Chapman (1935) for the maximum epoch of the main phase (to be defined shortly). It should be realized that this is an idealized average current system and that the instantaneous current system, derived from simultaneous measurement taken at stations distributed over the polar regions, is extremely variable and generally far more patchy than that of Figure 9 (Fukushima 1959, Fairfield 1963) although a general resemblance to the Ds current system of Chapman usually persists.

Figure 10 shows after Sugiura and Chapman (1960) the Dst variation and the amplitude of the main Fourier component (the diurnal component) of the Ds variation as a function of storm time for the first three days of weak, moderate and great storms, for a mean dipole latitude of 30° (the Ds amplitude is arbitrarily plotted as a negative quantity, for ease of comparison with Dst). It will be seen that Dst is positive for about the first three hours and then becomes negative. The positive Dst phase is usually called the initial phase and is ascribed (Chapman and Ferraro 1931, 1932) to the increased compression of the earth's magnetic field by charged particles emitted by the sun. Sometimes the arrival of this solar plasma is indicated by a sudden change in the magnetic field (a sudden commencement), which occurs almost simultaneously (within less than a minute) over the whole earth. The arrival of the solar plasma usually follows a day or two after an intense solar flare. It is to be noted that there is a continuous emission of particles from the sun (the solar wind); the particles responsible for the magnetic storm represent an additional emission.

The negative Dst phase, called the main phase, is usually ascribed to the adiabatic eastward drift of trapped energetic electrons and to the westward drift of trapped energetic protons (Singer 1957). Akasofu, Cain, and Chapman (1962) showed that the adiabatic drift of trapped protons observed by satellites during magnetically quiet days (Davis and Williamson, 1962) reduces the earth's magnetic field by about 38 γ at the equator. Presumably additional protons become trapped during a geomagnetic storm and reduce the earth's field still further until they are removed by charge exchange with neutral hydrogen; this could explain the main phase and the recovery phase of magnetic storms.

The above short discussion shows that reasonably good explanations of the Dst variations exist in terms of the increased initial compression of the magnetosphere by an intensified solar stream and in terms of a first gradually increasing and then decaying ring current due to additional trapped protons. In terms of these explanations residual currents corresponding to those responsible for the Dst variations would be expected to flow even during geomagnetically quiet days. Similar predictions are made by some of the theories of the Ds current system discussed in the next section. Analysis of quiet day magnetic data appears to show the presence of two residual current systems similar to those responsible for the Dst (Price 1963) and Ds (Nagata and Kokubun, 1962) variations respectively.

5.2 Theories of the Ds Current System

Since both the Ds and the Dst variations tend to be associated with geomagnetic storms, the question may be asked whether increased compression of the geomagnetic field by the solar stream and the presence of additional trapped protons could also play a role in the explanation of the Ds variations.

This was not the tendency in early explanations of the Ds variations. A modified tidal dynamo theory, in which an enhancement of ionospheric conductivity in the auroral regions was assumed, has been suggested by Obayashi and Jacobs (1957). H. Maeda (1957) and Swift (1963a, 1963b) have shown that the current system resulting from such an assumption does not resemble the Ds current system and that a very improbably atmospheric wind system has to be assumed in the auroral zone to explain the observed Ds variations.

In some of the more recent theories of the Ds variations the trapped particles of the "radiation belts" play an increasingly important part. Chamberlain et. al. (1960) have suggested that the auroral electrojets are caused by the precipitation of previously trapped energetic particles of different signs at slightly different latitudes. Chamberlain (1961), Kern (1961), and Fejer (1961) have independently suggested that charge separation, caused by the adiabatic motion of energetic trapped particles (given an asymmetric initial distribution), could lead to the auroral electrojet currents. This mechanism which invokes charge separation alone, tends to lead to only a temporary flow of currents until neutrality is restored in the belt by the adiabatic motion itself. If, however, the charge separation is followed by precipitation as in the mechanism of Chamberlain et. al. (1960), then a steady current system can result.

A steady current system can also be produced in a different manner, by the streaming of low energy magnetospheric plasma across a belt of energetic charged particles of predominantly one sign (Fejer 1963a, 1963b). The motion of the magnetospheric plasma may be partly caused by electric fields associated with the tidal dynamo currents but is probably mainly due to the participation of magnetospheric plasma in the earth's rotation. As will be shown in the next section, this participation is modified by geomagnetic distortion due to the solar wind in a manner that leads to a current system of the Ds type. This mechanism explains the Ds variations and the Dst variations in terms of the same physical phenomena; the distortion of the geomagnetic field by the solar stream and the presence of energetic trapped protons. There is however a difference: the two phenomena counteract each other in producing the Dst variations whereas they act in combination (multiplicatively) in producing the Ds variations. This could at least partly explain the rather different type of storm time variation of Dst and of Ds amplitude, shown by Figure 10; another difference is the requirement of sufficient ionospheric conductivity for the Ds current system but not for the Dst currents.

Axford and Hines (1961), in their discussion of magnetospheric motions, suggest a rather different explanation of the Ds current system. They maintain that these currents are caused by motions impressed by viscous interaction on the magnetospheric gas by the solar wind; above about 150 km both ions and electrons take part in magnetospheric motions but in the 90-150 km height region the ions are stopped by collisions with neutral particles and therefore the participation of the electrons constitutes a current.

It is not easy to assess the relative importance of these different mechanisms in the production of the Ds variations. The dynamo current system must undoubtedly be modified by the increased ionospheric conductivity produced by auroral zone particle bombardment and such a modification must make a contribution to the Ds current system. It appears, however from H. Maeda's (1957) work that the contribution must be a relatively small one.

It is difficult to assess the importance of magnetospheric motions driven by viscous interaction (Axford and Hines, 1961) without an investigation of the nature of these interactions. It should be remarked here that other mechanisms (Fejer 1963a, 1963b) suggested for the explanation of the Ds currents necessarily also lead to magnetospheric motions.

Charge separation caused by the adiabatic motion of trapped particles (Chamberlain 1961, Fejer 1961, Kern 1961) could make a significant contribution to the Ds current system whenever temporary asymmetries occured in the belt of trapped particles. The frequency of such occurences cannot be easily estimated. It is even more difficult to estimate the ionospheric currents caused by direct particle bombardment (Chamberlain et. al., 1960) since not enough is known about the nature of auroral bombardment.

The importance of ionospheric current generation by the interaction of magnetospheric motions with the belt of trapped particles can be estimated roughly from satellite data on trapped energetic protons and on the distortion of the geomagnetic field. These rough estimates tend to show (Fejer 1963a, 1963b) that the resulting current system could explain the small enhancement of magnetic variations in the auroral zone (Nagata and Kokubun 1962) during quiet days; the greatly enhanced magnetic variations during disturbed days would then have to be explained by increased geomagnetic compression and a greater temporary trapped proton population. The methods used in estimating the current system due to the interaction of magnetospheric motions with the belt of energetic protons will be shortly outlined here.

5.3 Currents due to the Presence of Energetic Trapped Protons in the Distorted Geomagnetic Field.

It is convenient to artifically divide the charged particles of the magnetosphere into two groups: 1) energetic particles with energies greater than say, 40 keV, whose orbits are hardly affected by the convection of the magnetosphere (i.e. by the presence of electrostatic fields there) because their drift motion due to the inhomogeneity of the magnetic field (westward for protons and eastward for electrons) is much faster than the velocity of magnetospheric convection and 2) low energy particles whose only motion is that due to the convection of the magnetosphere.

Consider the convection of the magnetosphere due to the rotation of the earth and the interaction of this rotation with a belt of energetic trapped protons. When the geomagnetic field is disturbed by the solar wind, the path of the equatorial crossing point of an energetic trapped particle is also distorted; the crossing points lie closer to the earth on the night side than on the day side. This is illustrated schematically by Figure 11, where the shaded region represents the equatorial crossing points of the proton belt in the distorted geomagnetic field. Figure 11 also shows a flow line of the rotating low energy plasma in the equatorial plane. As a line of force approaches the subsolar position, it reaches its maximum compression; thus the rotating low energy plasma is closest to the earth in the subsolar direction. The low energy plasma therefore tries to stream across the high energy proton belt, inward on the morning side and outward on the evening side, as seen from Figure 11. Such a motion tends to upset the balance of space charge since the excess negative space charge of the low energy particles that compensates for the positive space charge of the energetic protons, streams inward or outward. Charge neutrality is restored approximately by currents along the highly conducting field lines which link the magnetosphere to the ionosphere and by ionospheric currents. The ionospheric currents are driven by electric polarization fields due to the residual space charge. The electric fields are mainly in the north-south direction in the ionosphere. In addition to the "neutralizing" ionospheric currents whose direction is that of the driving electric field, there will also be intense ionospheric Hall currents which are approximately free of divergence and whose direction of flow is normal to the driving electric field. They are identified with the auroral electrojets.

It is to be noted that these electric fields produce drift motions in the magnetosphere which must be added to the motions caused by the earth's rotation. If the ionospheric conductivity is sufficiently high then these motions may be neglected; otherwise they constitute a reaction on the originally assumed motions which must be taken into account.

Since the motions of the magnetosphere have the nature of an interchange of tubes of force, it is convenient to consider the projection of the motions to the ionospheric shell along the lines of force which are approximately also lines of equal potential. Let a semistationary state be considered from a system that does not take part in the earth's rotation (as seen from the sun) and in which the density of trapped protons is only a function of position and does not vary with time. For simplicity both the rotational and the magnetic axes are taken to be normal to the ecliptic plane. It is convenient to assume symmetry about the equator and define the space charge-Q of energetic trapped protons per unit area on the ionospheric shell, after projection along the lines of force. If the geomagnetic field is distorted by the solar wind, then Q is not only a function of the latitude but also of the longitude (or more correctly of local time). The space charge per unit area of the low energy particles will then be approximately Q and this space charge moves with the projected velocity $\underset{\sim}{v}_R + \underset{\sim}{v}_D$ where $\underset{\sim}{v}_R$ is the projected velocity due to the earth's rotation and $\underset{\sim}{v}_D$ is the additional projected drift velocity caused by the presence of electric polarization fields seen from the rotating system. (It is assumed that the geomagnetic distortion due to the solar wind is negligible at ionospheric heights so that, apart from the additional drift velocity $\underset{\sim}{v}_D$, the feet of the lines of force in the ionosphere simply participate in the earth's rotation). In a stationary state there can be no steady accumulation of charges and therefore

$$\mathrm{div}\left(\int \underset{\sim}{J}_h \, dz + Q\,\underset{\sim}{v}_R + Q\,\underset{\sim}{v}_D\right) = 0 \tag{22}$$

where $\int \underset{\sim}{J}_h \, dz$ is the height integrated ionospheric current density and $\underset{\sim}{J}_h$ is given by equation (20). The south and east components of v_R are

$$v_{Rx} = 0 \qquad , \qquad v_{Ry} = \omega R \sin\theta \tag{23}$$

where R is the radius of the ionospheric shell. The drift velocity v_D of the foot of a line of force has the components

$$v_{Dx} = (\partial\Psi/\partial y)/(B_0 \sin I) \qquad , \qquad v_{Dy} = -(\partial\Psi/\partial x)/(B_0 \sin I) \tag{24}$$

A combination of equations (20) and (22) yields the differential equation

$$\mathrm{div}\left[\int \underset{\sim}{\sigma}_h \cdot (\underset{\sim}{u} \times \underset{\sim}{B}_0)_h \, dz - \mathrm{grad}_h \Psi \cdot \int \underset{\sim}{\sigma}_h \, dz + Q(\underset{\sim}{v}_R + \underset{\sim}{v}_D)\right] = 0 \tag{25}$$

for Ψ, of which equation (21) of the dynamo theory is a special case.

Equation (25) was solved numerically by Fejer (1963b). In his analysis $\underset{\sim}{u}$ was set equal to zero. The same ratios of the height integrated conductivities $\int \sigma_0 dz$, $\int \sigma_1 dz$ and $\int \sigma_2 dz$ were used as in Fejer's (1953) previous work on the dynamo theory. These height integrated conductivities were assumed to be independent of geographical position; the resulting components of $\int \underset{\approx}{\sigma}_h dz$ were thus functions of the latitude only. The latitude distribution of the function Q was estimated from the satellite observations of Davis and Williamson (1962) on trapped energetic protons. An artifical longitudinal asymmetry was then introduced which shifted the latitude distribution of Q nearer to the pole by 7° on the day side, and further from the pole by 7° on the night side, in order to simulate roughly the geomagnetic distortion due to the solar wind. The corresponding maximum projected streaming velocity of low energy plasma across the proton belt would be about 58 m/sec.

The computed ionospheric current systems are shown by Figure 12a-12c for three different sets of values of the height integrated conductivities. The electric fields driving the currents may be derived from Figures 13a-13c, which show lines of equal potential on the ionospheric shell. The Hall currents flow parallel to the lines of equal potential which can therefore also be interpreted as lines of flow of the Hall current. Under the present assumptions, the Hall current is not very different from the total ionospheric current except near the equator.

Figures 12a and 13a show that the current system closely resembles Chapman's (1935) idealized Ds current system (Fig. 9) for $\int \sigma_2 dz$ = 47 mhos (about twice the midday conductivity at medium latitudes) and that a height-integrated maximum east-west current density of 63 amps/km (corresponding to a change in the horizontal component of the magnetic field of about 80 γ) flows in the auroral zone. This current density is not decreased substantially when the conductivity is dropped to $\int \sigma_2 dz$ = 14 mhos but the shape of the current system is changed as shown by Figures 12b and 13b. As the conductivity is decreased still further to $\int \sigma_2 dz$ = 46 mhos, the maximum east-west current density in the auroral zone is reduced to about 22 amps/km and the shape of the current system is modified considerably as shown by Figures 12 c and 13c, but the magnitudes of the current densities in the polar caps and at low latitudes are not yet modified substantially. As the conductivity is further reduced, the calculations show a reduction of the current density at all latitudes.

The results of these calculations show that at geomagnetically quiet times the interaction of magnetospheric motions with the belt of trapped protons could lead to changes in the horizontal field by about 70 γ in the auroral zone on the day side; the necessary ionospheric conductivity is then produced by ultraviolet radiation from the sun. It is equally clear from these calculations that the explanation of observed changes of several hundred γ in the horizontal component of the magnetic field in the auroral zone on the night side during magnetic storms requires not only more trapped protons and a greater geomagnetic distortion but also auroral zone night-time ionospheric conductivities which exceed by at least a factor of about 2 the midday conductivity at moderate latitudes. Such high conductivities probably occur in the auroral zone during severe sporadic E conditions. The patchiness of the actual current distribution may well be caused by the patchiness in the ionospheric conductivity.

Fejer (1963b) shows that if u is not set equal to zero then the presence of trapped energetic protons modifies the dynamo current system even in the absence of any distortion of the geomagnetic field by the solar wind. The additional current system resembles those of Fig. 12 and 13 in shape but has the opposite phase if the generally accepted phase of the tidal polarization field (H. Maeda 1955) is assumed. Thus the mere presence of energetic trapped protons could produce considerable additional magnetic variations in the auroral zone. The actual current system probably partly consists of this modified dynamo current system and is partly due to geomagnetic distortion by the solar wind.

5.4 Diurnally Recurring Events. Magnetically Conjugate Events.

It was shown in the previous section that the projection of the belt of trapped particles to the earth's surface along the field lines of the distorted geomagnetic field is not symmetrical about the geomagnetic pole. If a narrow shell of trapped particles is considered then the projection is at a higher latitude near the subsolar point than on the night side. From a system rotating with the earth this asymmetric projection of the shell rotates about the magnetic axis with a period of one solar day so that a given point on the earth may find itself on the projection twice a day at two separate local times. Since the magnetic effects are strongest near the projection of the shell, the observed daily recurrence of certain features of magnetic records for several successive days (Chapman and Bartels 1940) may be explained in this manner. The striking similarity of magnetic records at conjugate stations (Wescott 1961, 1962) may sometimes have similar causes although simultaneous changes in the precipitation of particles from the common field line and the resulting simultaneous changes in ionospheric conductivity may play an even greater part.

6. Additional Effects Associated with Ionospheric Current Systems

6.1. Magnetospheric Motions

It was pointed out previously that the ionospheric current system due to the presence of energetic trapped protons in the distorted geomagnetic field (as well as the tidal dynamo current system) is accompanied by magnetospheric motions. These motions may be derived from Figures 13a-13c since the lines of equipotential given there are also flow lines of the projected motion. Since the velocity of the motion is proportional to the electric field, it follows from Fig. 13a-13c that the velocity of the motion reaches it maximum value for low conductivities and decreases to very low values with increasing conductivity. The convection pattern is similar to that suggested by Axford and Hines (1961) and will not be discussed here further. It should however be mentioned that moving auroral radar echo patterns, moving visual auroral forms and drifting ionospheric irregularities responsible for radio star scintillations could all be regarded as observational evidence for the existence of magnetospheric motions.

A less direct evidence of magnetospheric drift motions associated with the tidal dynamo current system is the distortion of ionospheric layers which has been attributed to vertical drifts of ionization (Martyn 1950, K. Maeda 1955).

6.2. Ionospheric Heating Effects

The ionospheric currents responsible for magnetic variations are associated with Ohmic losses, and therefore cause atmospheric heating. (Cole 1962, Kato 1962). The heating in the 130 - 200 km region where the Pedersen conductivity is considerable, is likely to be particularly important. Heating effects of this nature would be expected to be most intense in the auroral zone where the electric fields are largest. Further analyses of these heating effects and their possible significance in the explanation of the increased atmospheric drag observed by satellites during geomagnetic storms (Jacchia 1959a, 1959b) would be desirable.

Electric fields at heights of about 400 km have also been invoked (Megill, Rees and Droppelman, 1962; Megill and Carleton, 1963) in attempts to explain the night sky emission of the red lines of atomic oxygen in mid-latitude red arcs (Roach and Roach, 1963). The excitation is assumed to be caused by electrons whose temperature may exceed the temperature of the neutral particles considerably when an electric field normal to the magnetic field is present.

6.3. The Acceleration of Charged Particles

Two types of acceleration mechanisms could be associated with the current systems discussed here and particularly with the auroral Ds current system. The first of these has been proposed by Axford and Hines (1961). In their mechanism the acceleration is due to the compression suffered by a magnetospheric tube of force, as it moves from high to low latitudes; the converse process of deceleration is, of course, also possible. The acceleration may alternatively be described (Hines, 1963b) as being due to electric fields of the type depicted in Fig. 13 and is therefore limited at any given time by the available potential differences to tens of kilovolts.

A rather different type of acceleration mechanism may operate if strong inhomogeneities, such as a sharp latitudinal gradient, existed in the distribution of energetic trapped protons. Such inhomogeneities would lead to the sudden accumulation of space charges on the drifting tubes of force and these could not be conducted away instantaneously along the tubes of force as assumed in section 5.3. Large electric fields could then build up, at least temporarily, along the tubes of force and these would lead to the acceleration of plasma electrons. Similar acceleration mechanisms were discussed by Chamberlain (1961, 1962). They may play an important role in the auroral electron bombardment and their further investigation would be desirable.

7. Concluding Remarks

The review given in this paper is intended to serve as a basis for further investigations, both theoretical and observational, of the daily magnetic variations.

It appears that all daily variations of the geomagnetic field are caused by the rotation of the earth as seen from the sun or the moon and that the variations could therefore be properly called tidal if that adjective were used in its widest sense. Gravitational effects are believed to be entirely responsible for lunar tidal geomagnetic variations but they appear to play only a minor part in the generation of daily magnetic variations related to the solar day. The latter are probably caused partly by the atmospheric absorption of solar radiation mainly below the mesopause (this restriction is very tentative since the effects of a heat input above the mesopause have not been considered) over the sunlit part of the earth and partly by the interaction of the solar wind with the geomagnetic field and the particles trapped in it. The excitation of atmospheric oscillations by the atmospheric absorption of solar radiation is believed to be the predominant cause of daily variations during magnetically quiet days (the Sq variations) whereas the interaction of the solar wind with the magnetosphere is thought to be the cause of daily variations during disturbed days (the Ds variations).

The quiet day magnetic variations are better known observationally and are much better understood theoretically than the disturbance daily variations. The integrated theory of the atmospheric tidal oscillations and of their magnetic effects is, however, still an unsolved problem. Moreover many observers are beginning to realize that the same mechanisms that are responsible for the generation of additional daily magnetic variations during disturbed days, are never quite absent, even during magnetically quiet days.

Theories of the daily disturbance variations are of relatively recent origin. Their present review is unavoidably biased since the cause of these variations is not yet known with certainty. Observational checks of the proposed theories are urgently needed. Satellite observations will be very helpful in this respect but a great deal more could probably be learned from ground-based magnetic observations and from the analysis of existing magnetic records.

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CAPTIONS FOR FIGURES

Figure 1 Schematic illustration of the tide-producing force as the resultant of the gravitational and the inertial forces.

Figure 2 Comparison of the atmospheric temperature profile used by Weekes and Wilkes (1947) with that of the 1962 U. S. Standard Atmosphere.

Figure 3a Calculated height distribution of the relative amplitude of solar semidiurnal pressure oscillations. Curve (a), Small and Butler (1963); curve (b), Weekes and Wilkes (1947).

Figure 3b Calculated height distribution of the phase of solar semidiurnal pressure oscillations. The oscillation is described in terms of cos (σ t + phase) where t denotes solar local time beginning at noon. Curve (a), Small and Butler (1963); curve (b), Weekes and Wilkes (1947).

Figure 4 The height distribution of the three conductivities σ_0, σ_1 and σ_2 for representative values of the collision and gyro-frequencies at a moderate latitude. At the height z_e the angular gyro-frequency of an electron is equal to its collision frequency; z_i has the same significance for ions.

Figure 5 The height integrated midday conductivities σ_{xx}, σ_{xy} and σ_{yy} as functions of the latitude.

Figure 6 World-wide wind system (shown by arrows) produced by the progressive component of the world wide pressure system. The pressure system is indicated by isobaric curves. The pressure difference between neighboring isobars is 0.2 mm Hg. (After Bartels in Wien-Harms, "Handbuch der Experimental physik," Vol. 25, Leipzig, 1928).

Figure 7 The overhead current system corresponding to Sq over the sunlit hemisphere (a) and the night hemisphere (b), sunspot minimum, at equinox. A current of 10,000A flows between adjacent lines. From Chapman and Bartels, "Geomagnetism," Vols. 1 and 2, Oxford, Clarendon Press, 1940.

Figure 8 The diurnal (a) and semidiurnal (b) wind patterns of the northern hemisphere, as deduced by Kato (1956) from the average quiet day magnetic variations. The sides of each elementary square represent a wind speed of 50 m sec^{-1}.

Figure 9 Views from the sun (a) and from above the north pole (b) of the idealized overhead current system that could produce the Ds magnetic variations. A current of 10,000A flows between adjacent lines. From Chapman and Bartels, "Geomagnetism," Oxford, Clarendon Press, (1940).

Figure 10 A comparison of the rates of evolution of Dst and the range of Ds during the first three days of weak, moderate and great storms; mean dipole latitude 30°. Dst curves are drawn with full lines, and those for Ds with broken lines. (Sugiura and Chapman, 1960).

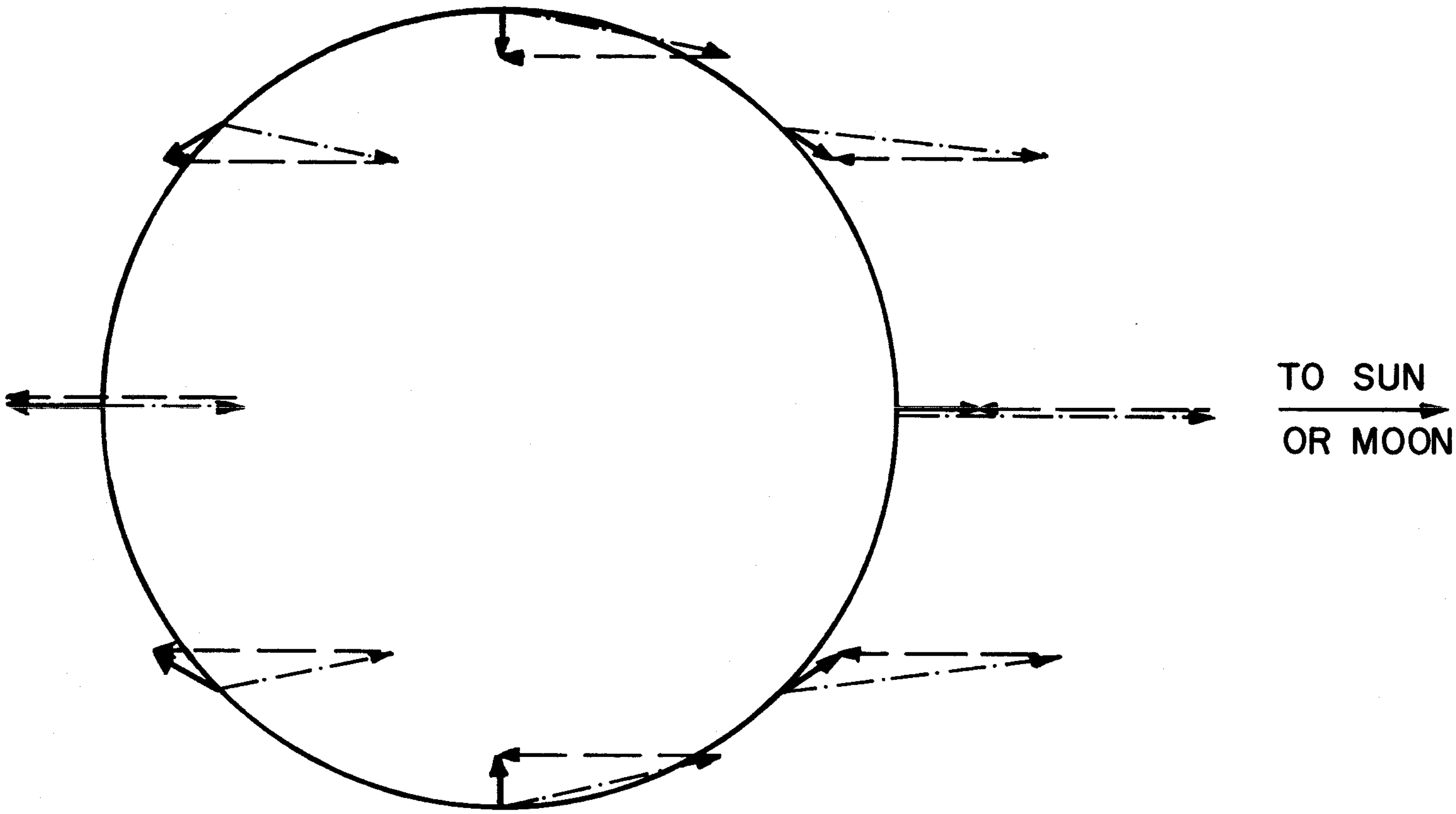
Figure 11 Schematic representation in the equatorial plane of the displaced belt of trapped protons and of a stream line of the low energy plasma rotating in the geomagnetic field distorted by the solar wind. The low energy plasma streams inward across the proton belt during the morning hours and outward during the evening hours.

Figure 12a-12c The calculated amplitudes and phases of the height-integrated current density components as functions of the latitude, for different values of the height-integrated Hall conductivity

$\int \sigma_2 dh$. The other conductivities were changed in proportion. The labels E and S denote the east and the south components of the height-integrated current density. The solid lines indicate the highest absolute value attained by each component during its diurnal variation; the interrupted lines show the local times when the sign of each component changes from positive to negative (east to west or south to north).

Figure 13a-13c Lines of equipotential of the electric polarization field driving the currents of Figure 12a-12c. The values of the height-integrated Hall conductivity $\int \sigma_2 dh$ and of the potential difference ΔV between neighboring lines are shown in each of the figures. The lines of equipotential can also be interpreted as lines of Hall current flow; the arrows indicate the direction of current flow.

FIG. I.

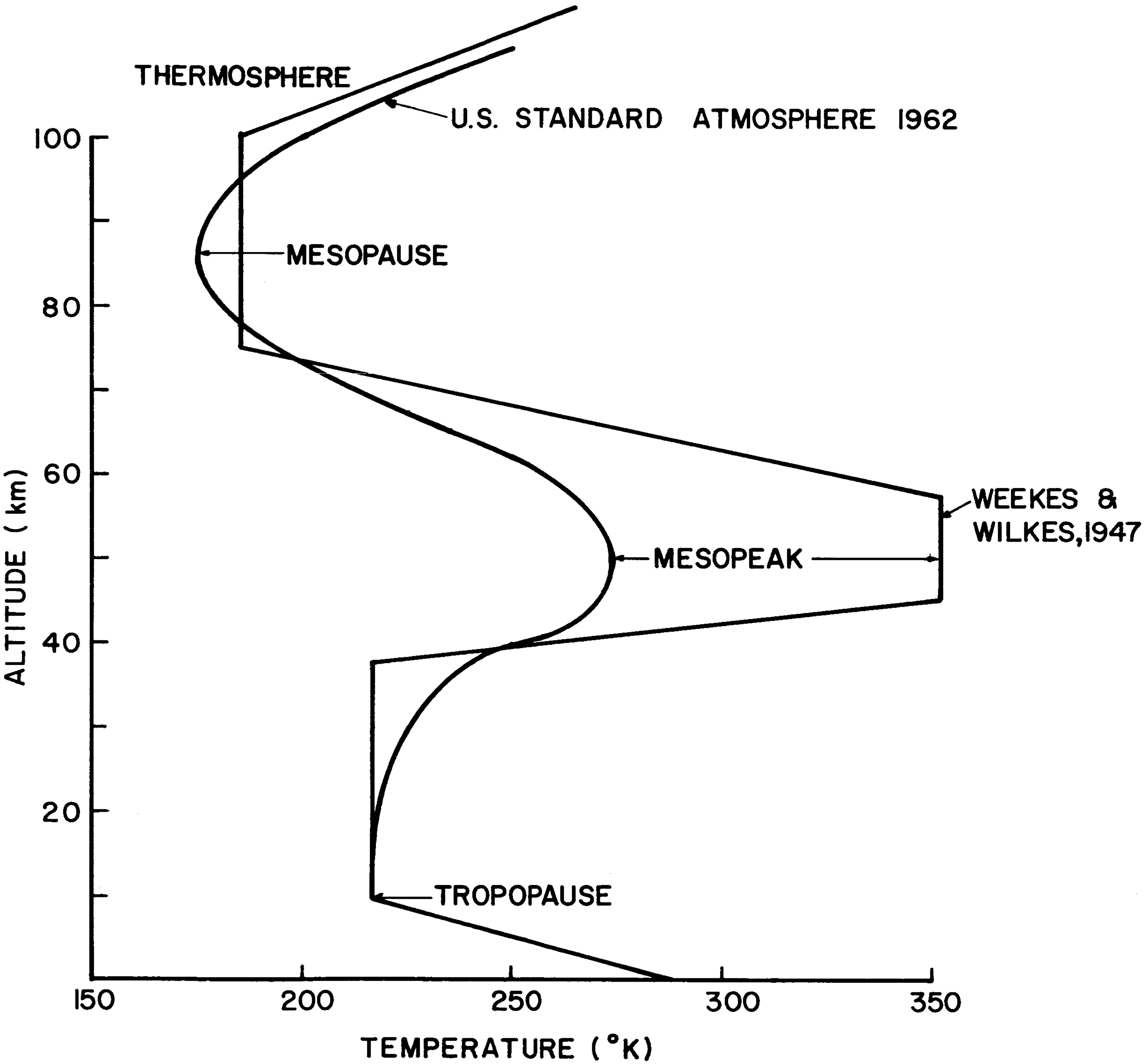


GRAVITATIONAL FORCE

INERTIAL FORCE

RESULTANT

FIG. 2.


THERMOSPHERE
U.S. STANDARD ATMOSPHERE 1962
MESOPAUSE
WEEKES & WILKES, 1947
MESOPEAK
TROPOPAUSE
ALTITUDE (km)
100
80
60
40
20
150
200
250
300
350
TEMPERATURE (°K)

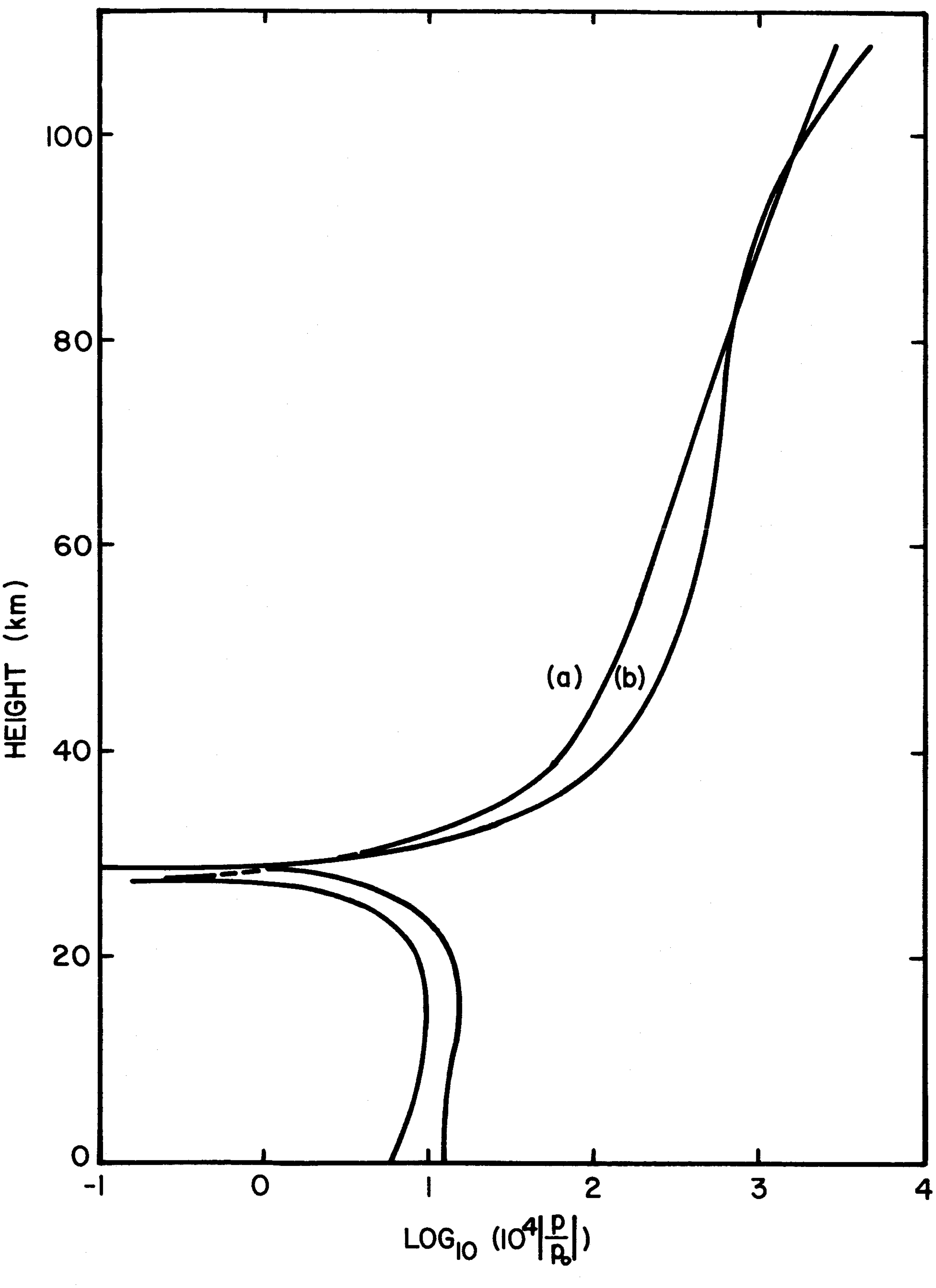

HEIGHT (km)
100
80
60
40
20
0
-1
0
1
2
3
4
(a)
(b)
$LOG_{10}\ (10^4 \left|\frac{p}{p_0}\right|)$

FIG. 3 a

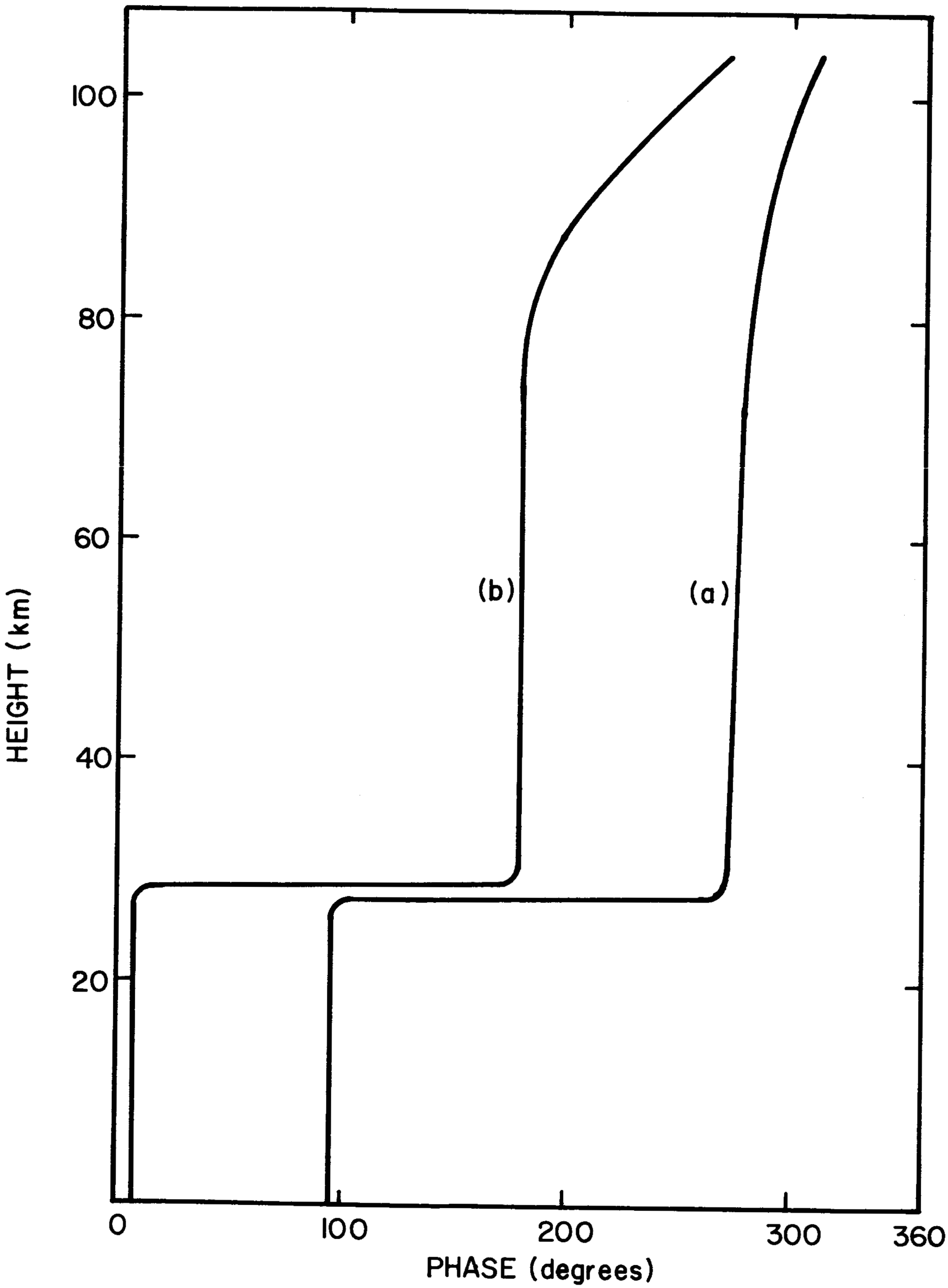


FIG. 3b

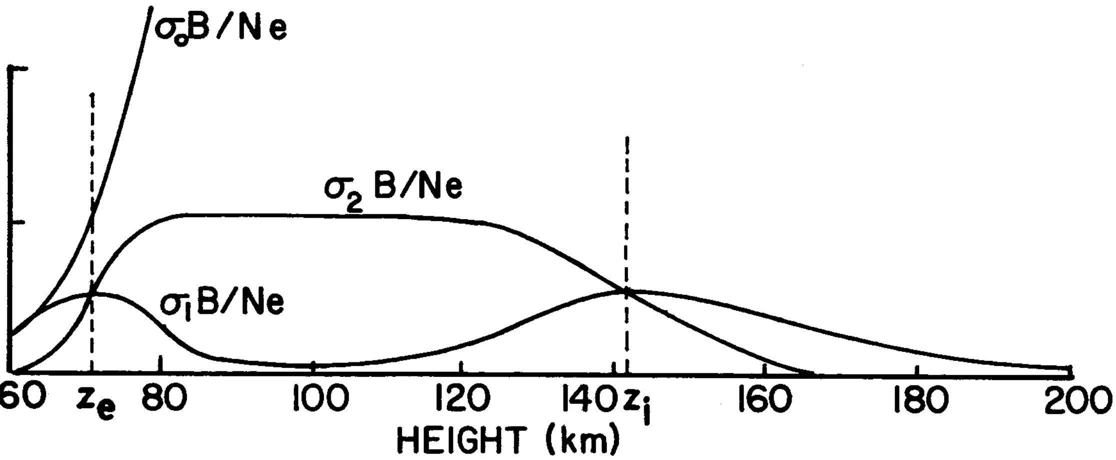


FIG. 4.

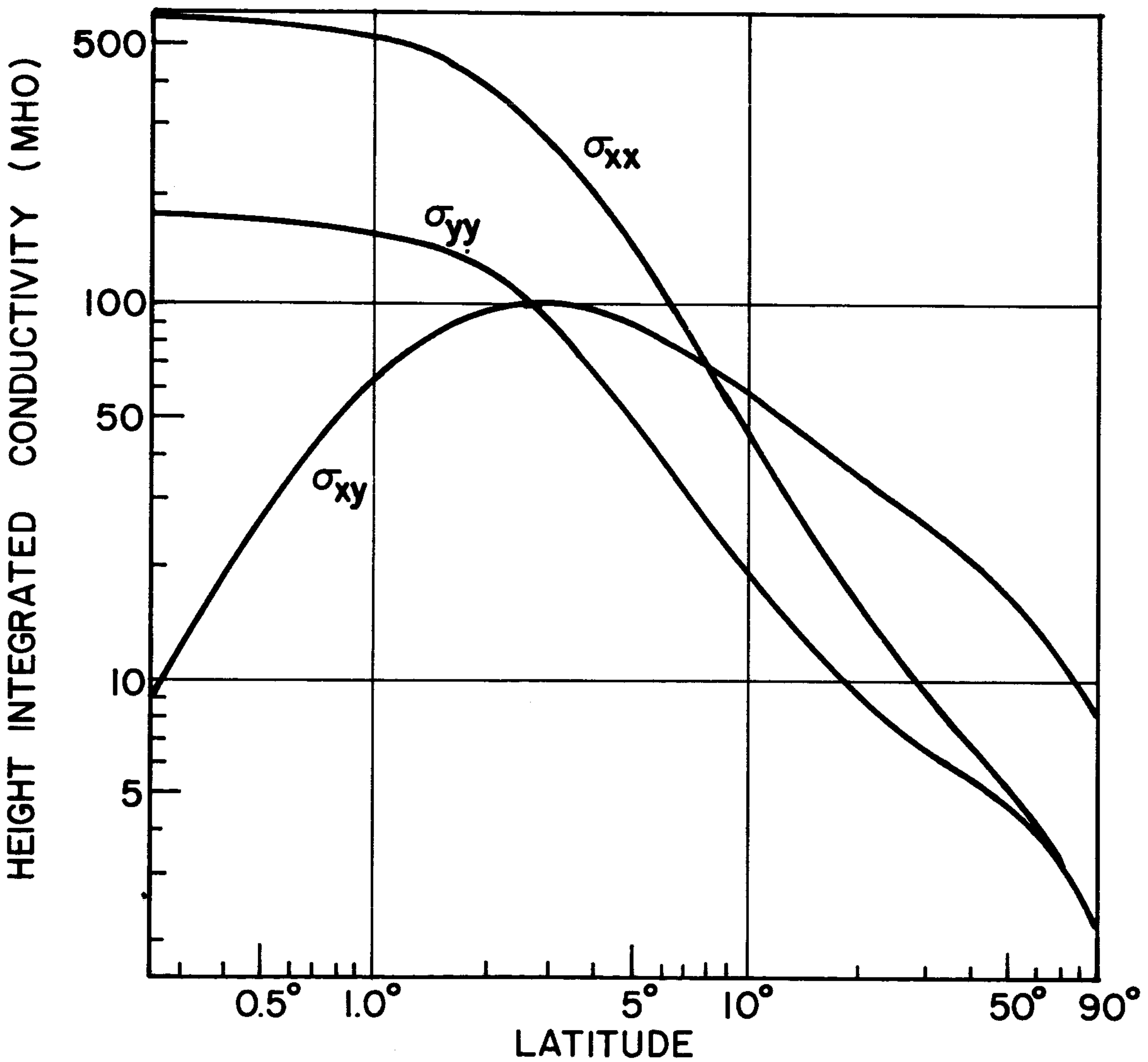


FIG. 5. THE HEIGHT INTEGRATED CONDUCTIVITIES σ_{xx}, σ_{xy}, AND σ_{yy} AS FUNCTIONS OF THE LATITUDES.

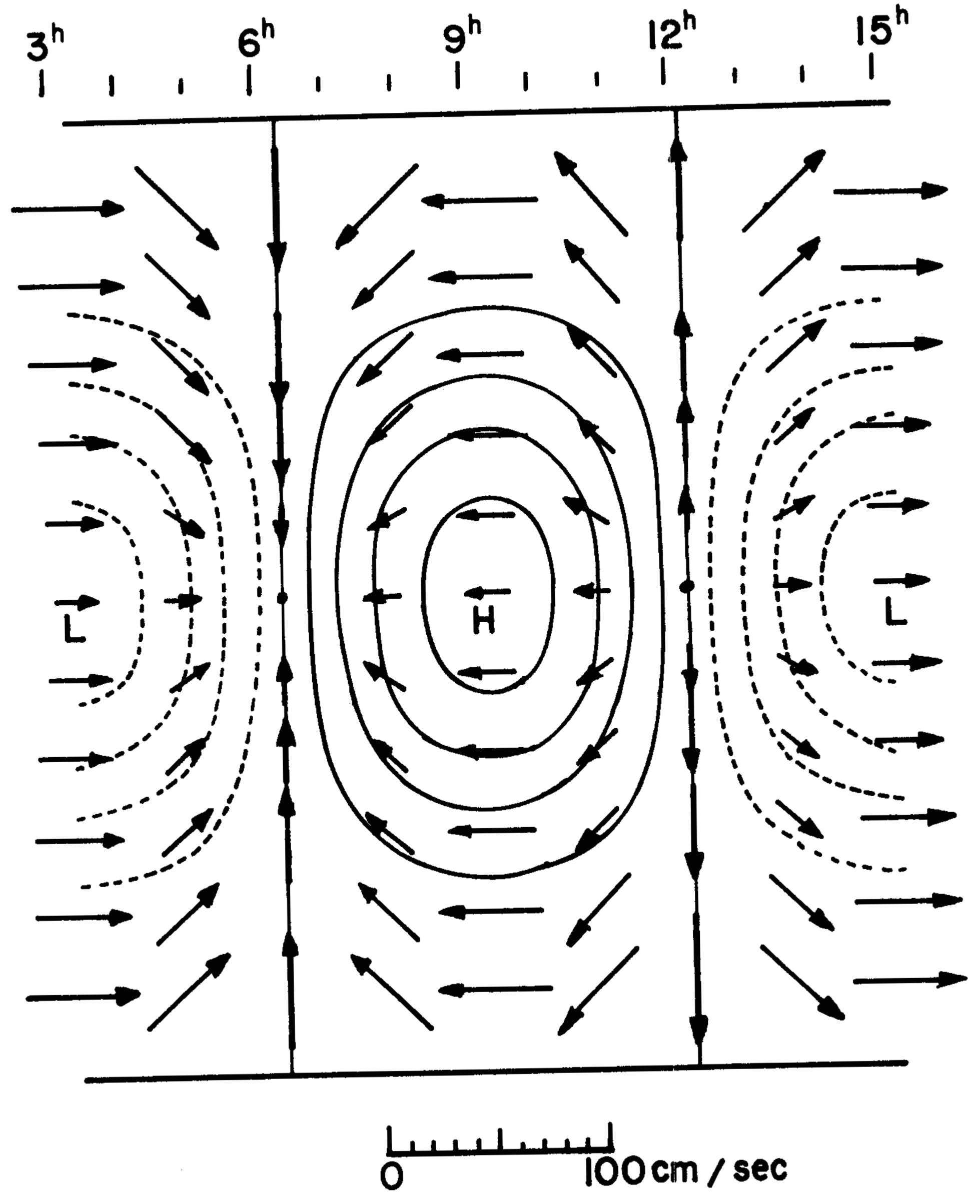
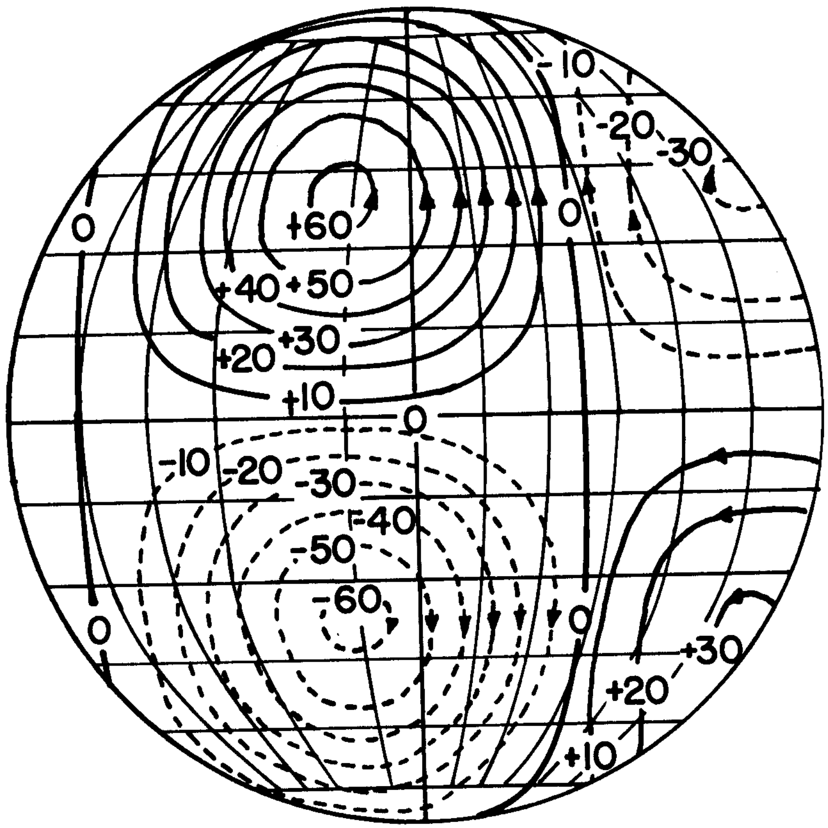


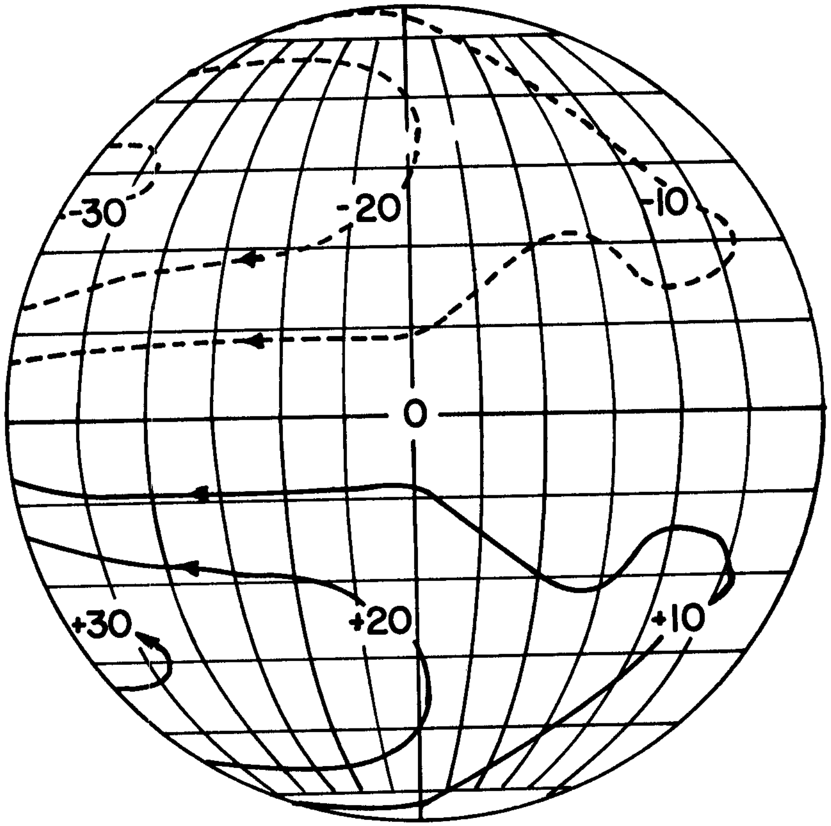
FIG. 6.

FIG. 7



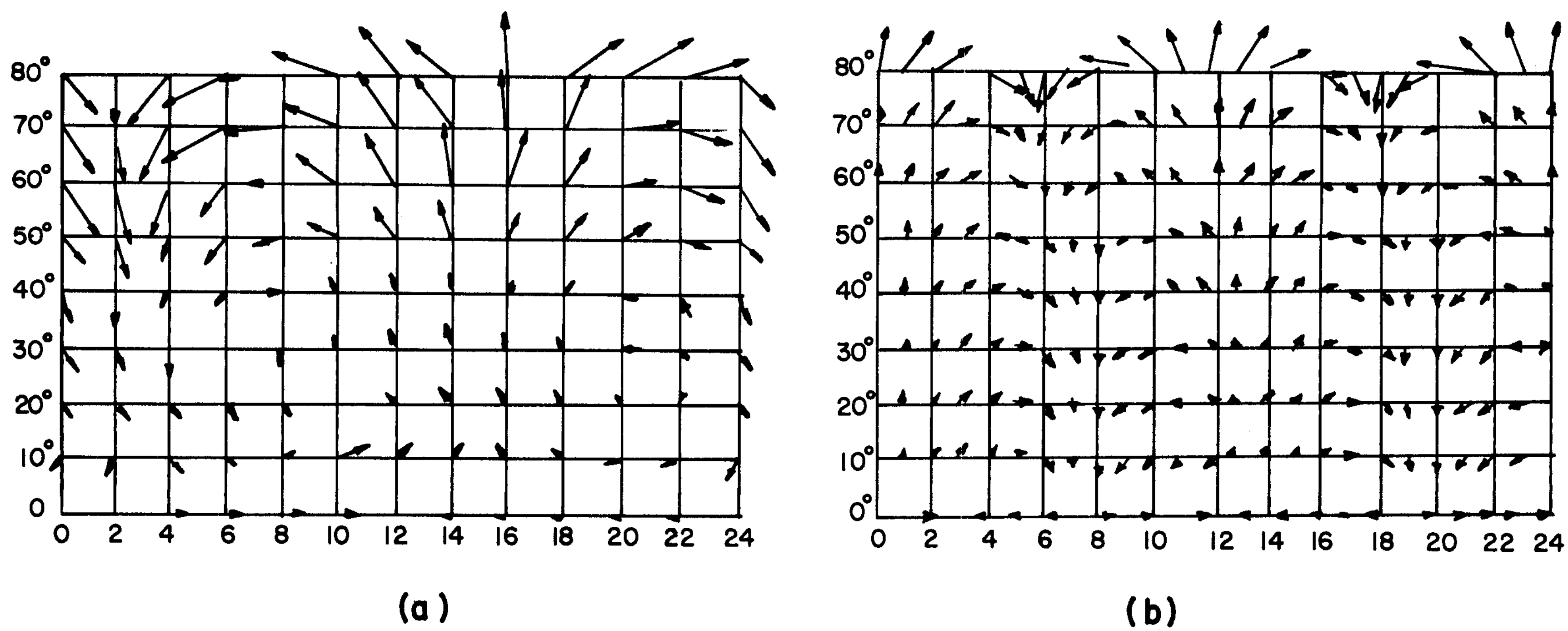
+60
+50
+40
+30
+20
+10
0
-10
-20
-30
-40
-50
-60
0
0
0
0
-10
-20
-30
+10
+20
+30

(a)



-30
-20
-10
0
+30
+20
+10

(b)



80°
70°
60°
50°
40°
30°
20°
10°
0
0 2 4 6 8 10 12 14 16 18 20 22 24
(a)
80°
70°
60°
50°
40°
30°
20°
10°
0°
0 2 4 6 8 10 12 14 16 18 20 22 24
(b)

FIG. 8.

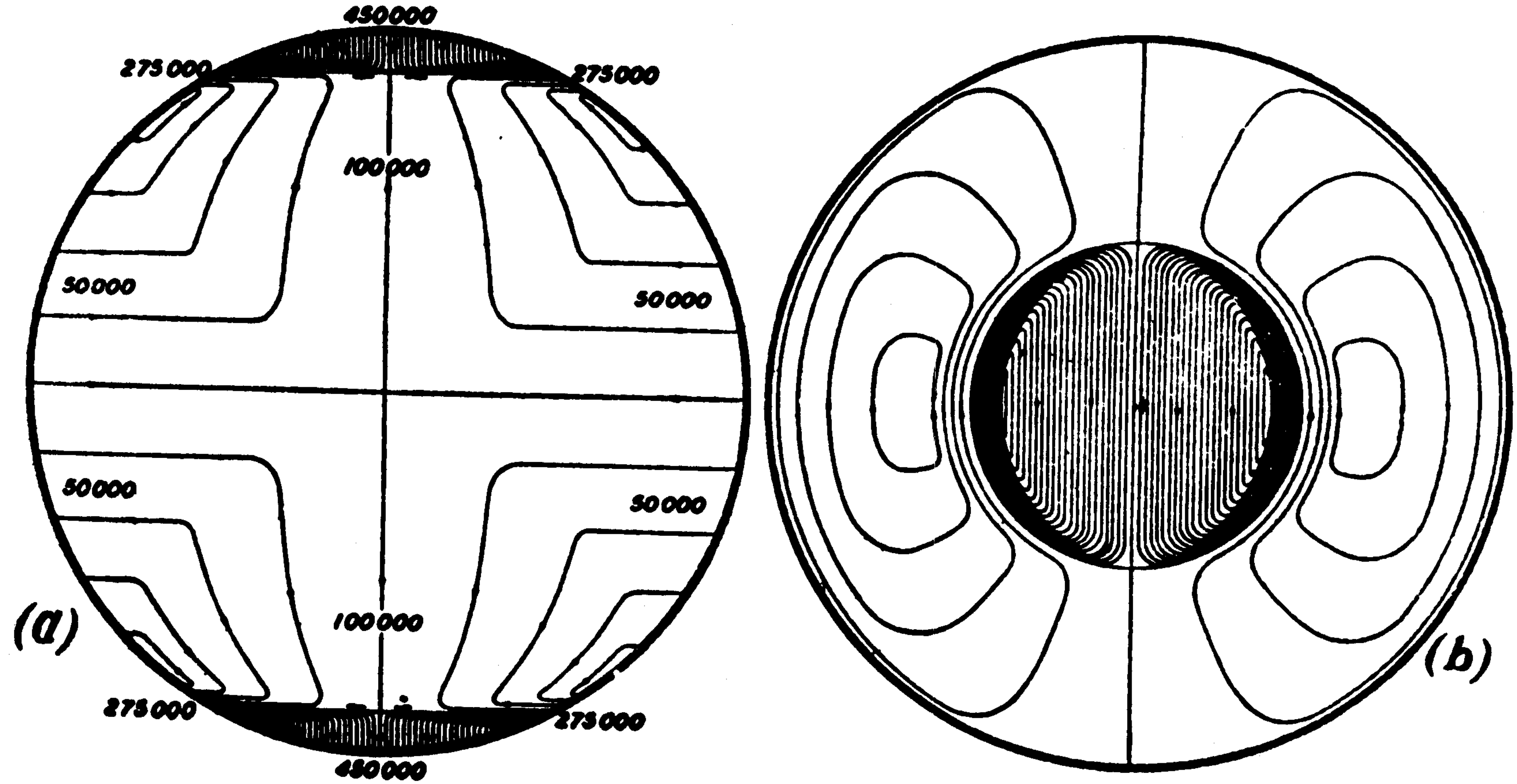

450000
275000
275000
100000
50000
50000
50000
50000
100000
275000
275000
450000
(a)
(b)

FIG. 9

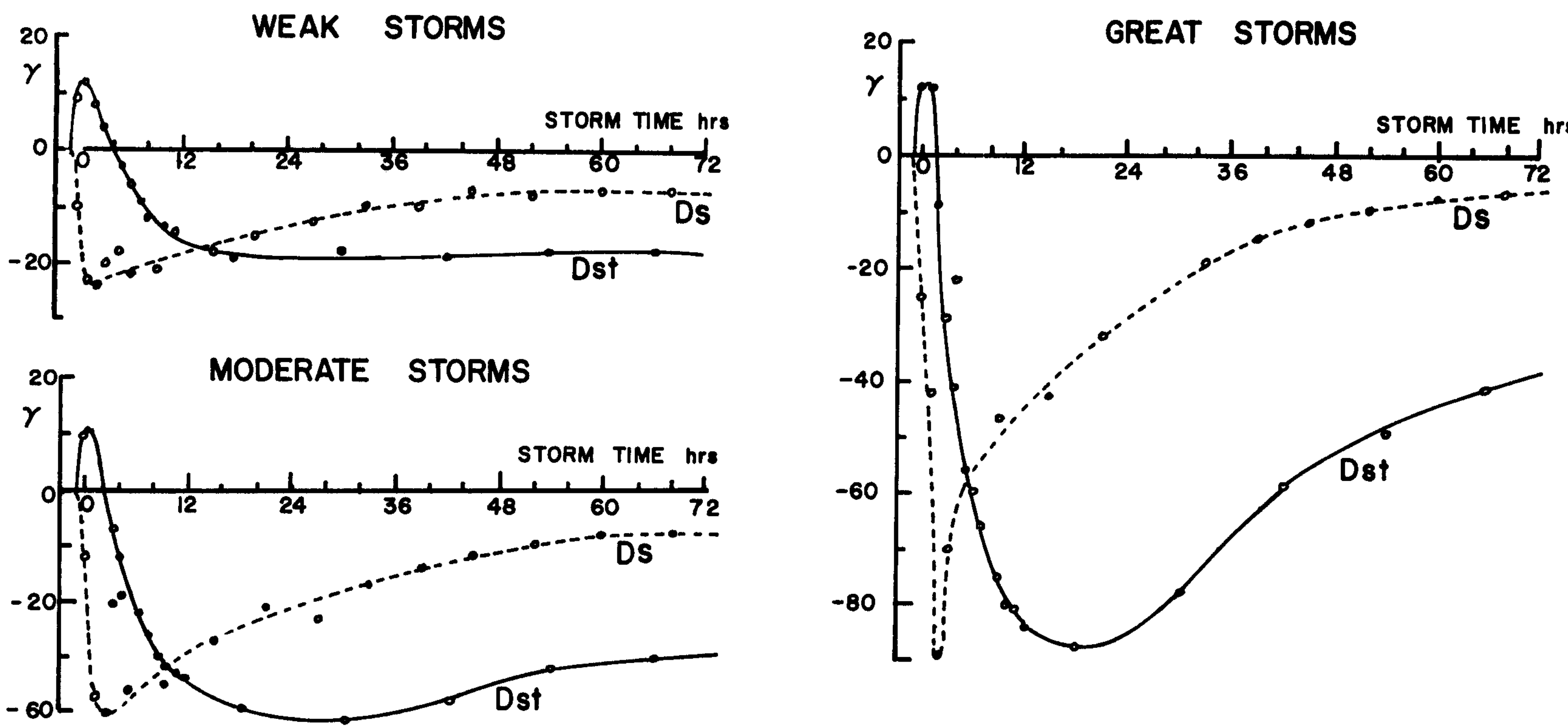

WEAK STORMS
MODERATE STORMS
GREAT STORMS
STORM TIME hrs
γ
Ds
Dst
20
0
-20
-40
-60
-80
0
12
24
36
48
60
72

FIG. 10.

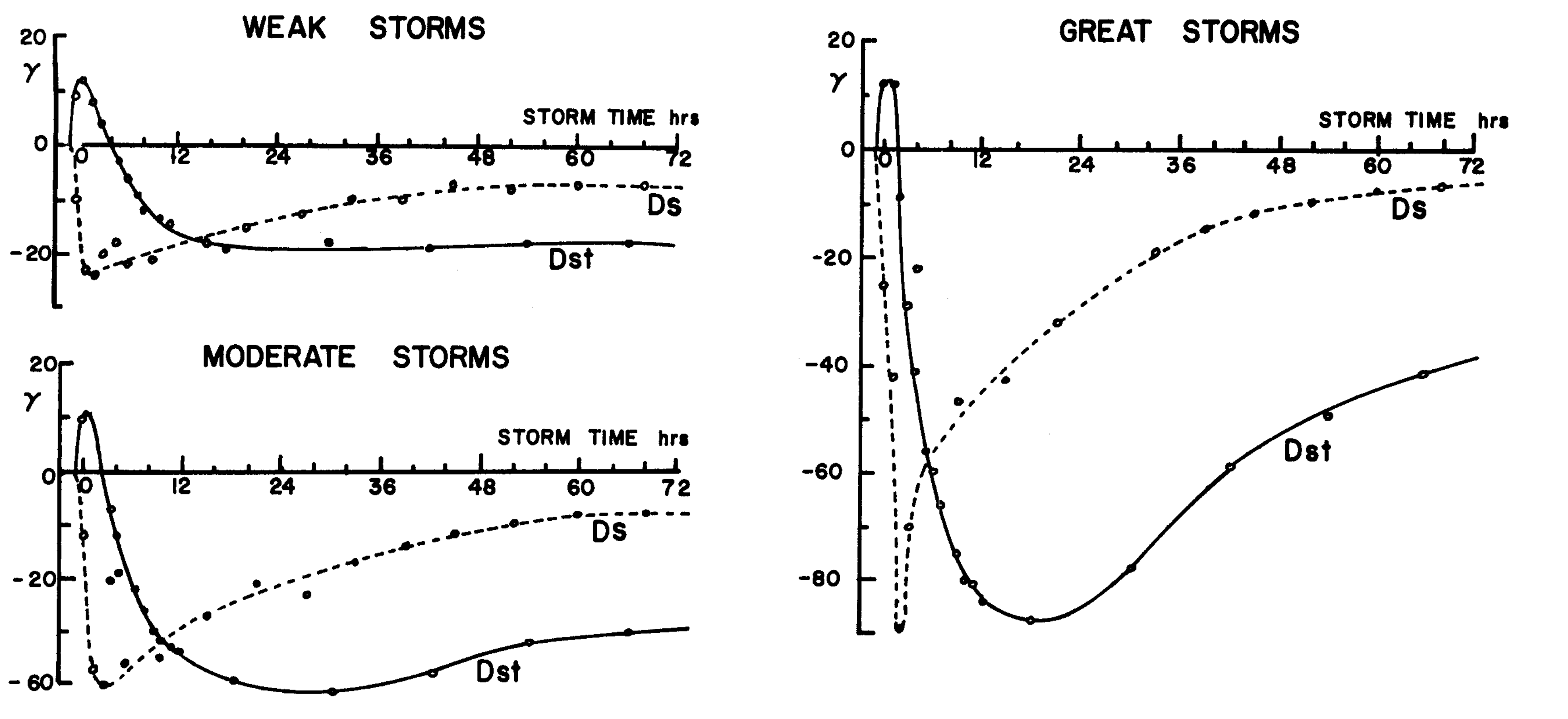
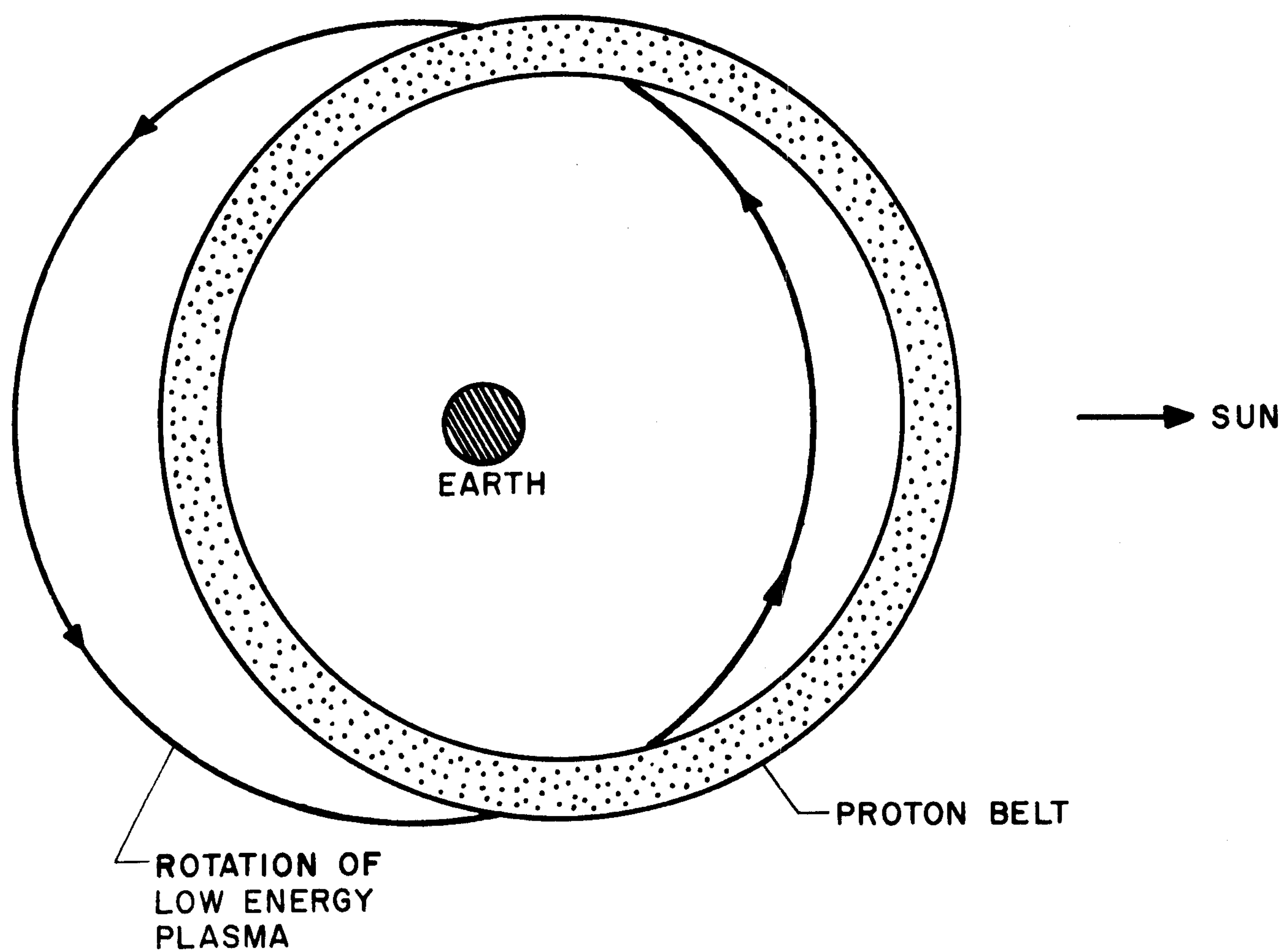
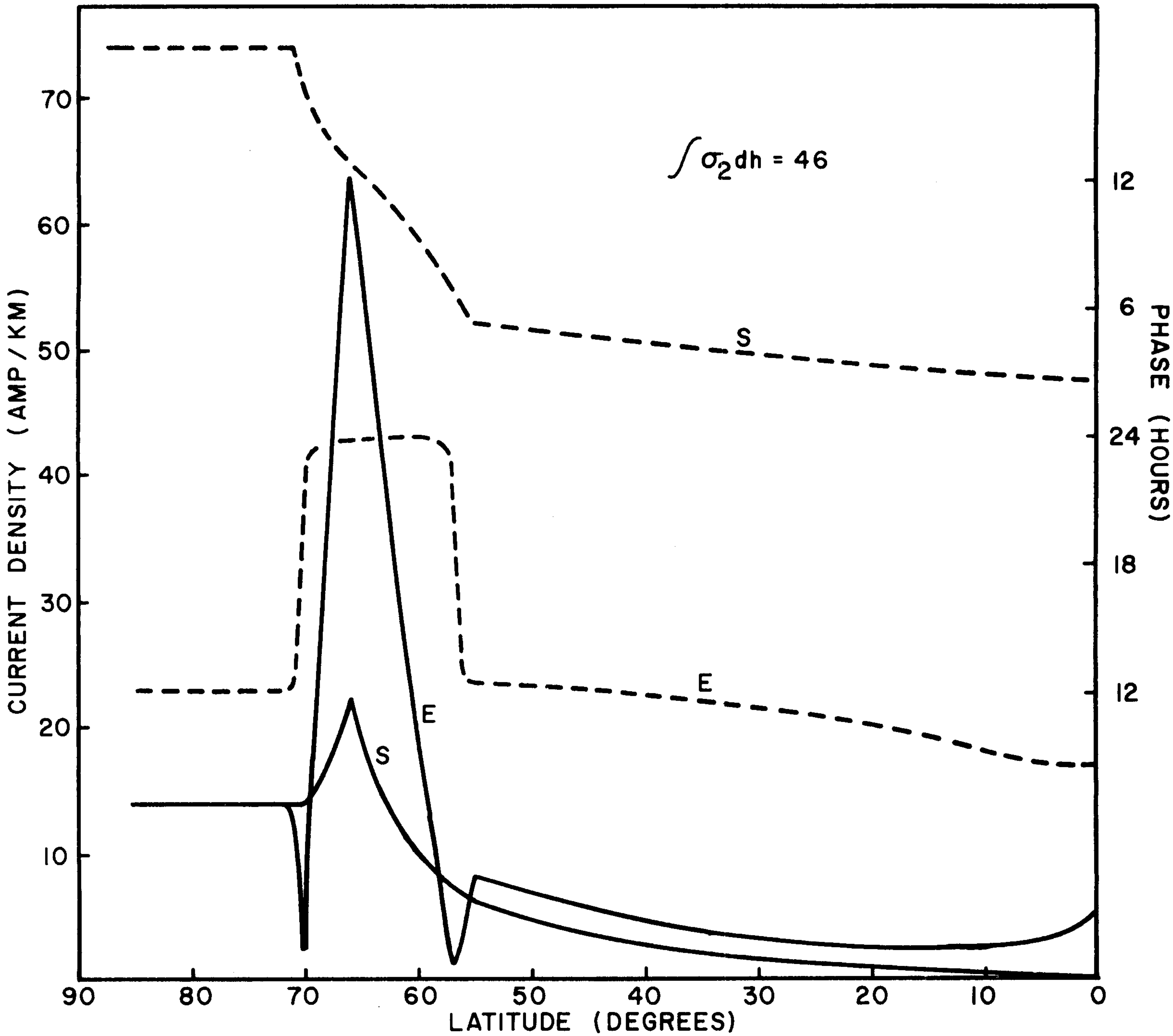
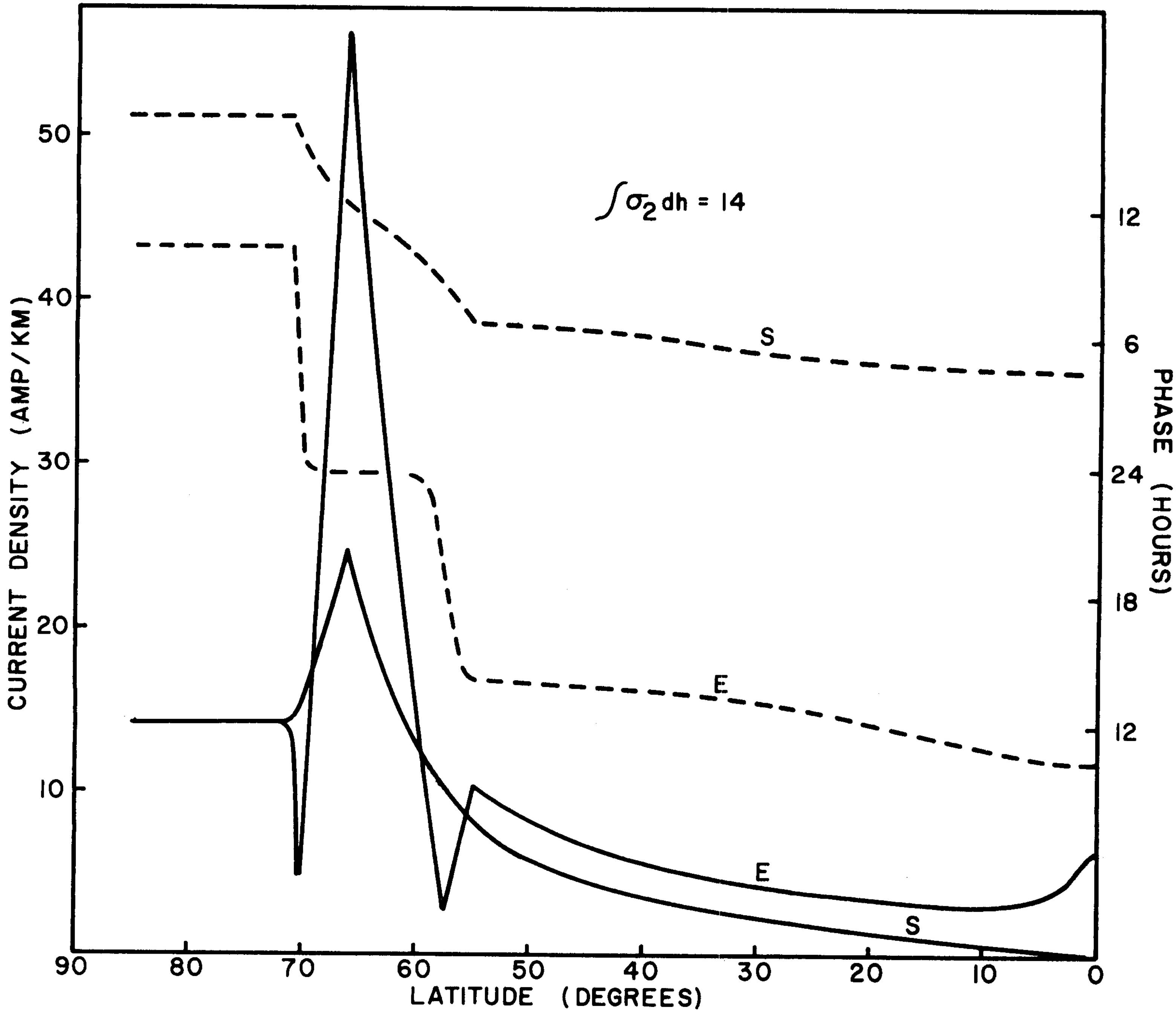

WEAK STORMS
MODERATE STORMS
GREAT STORMS
STORM TIME hrs
γ
Ds
Dst
20
0
-20
-40
-60
-80
0
12
24
36
48
60
72

FIG. 10.

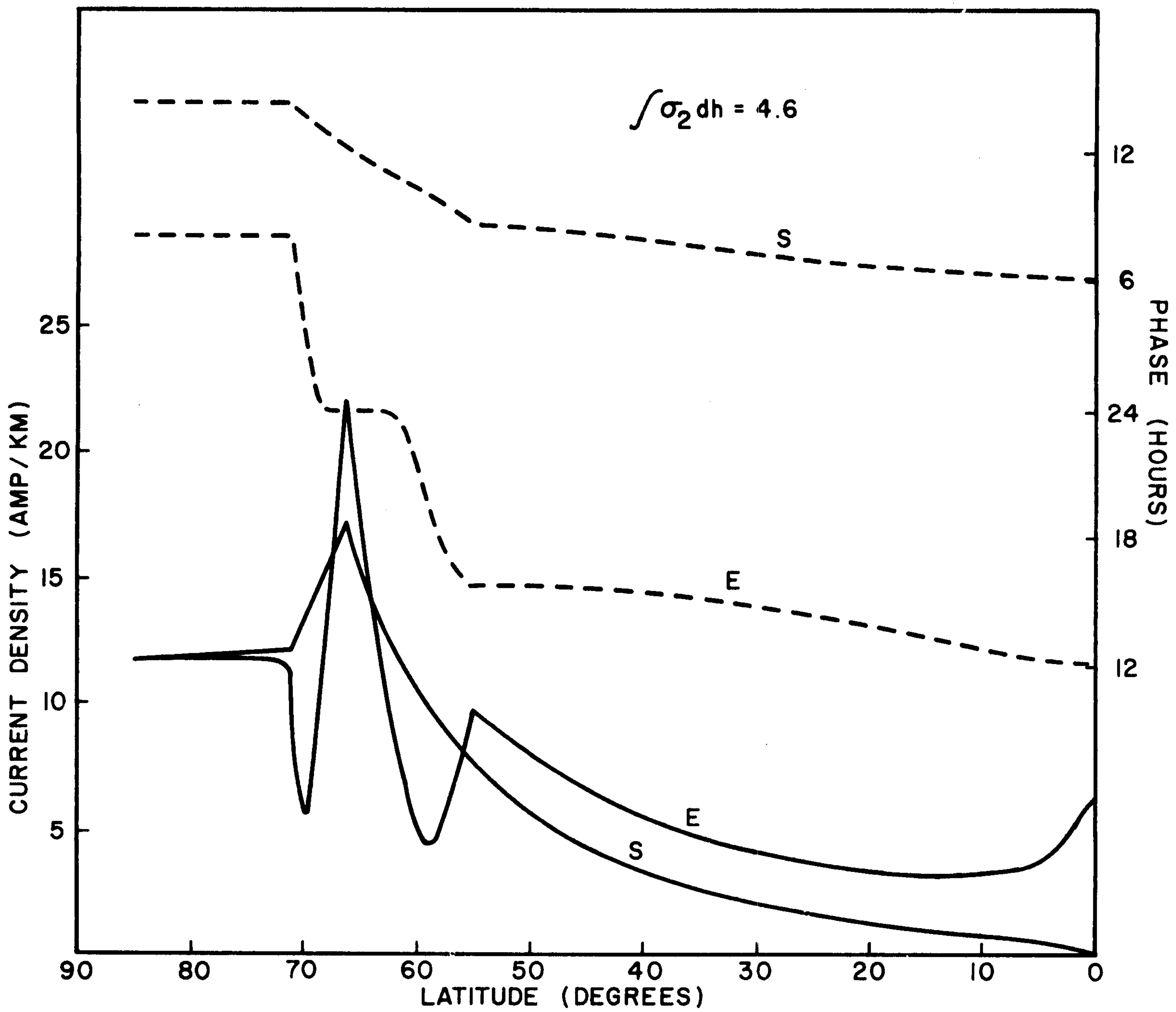

SUN
EARTH
PROTON BELT
ROTATION OF
LOW ENERGY
PLASMA



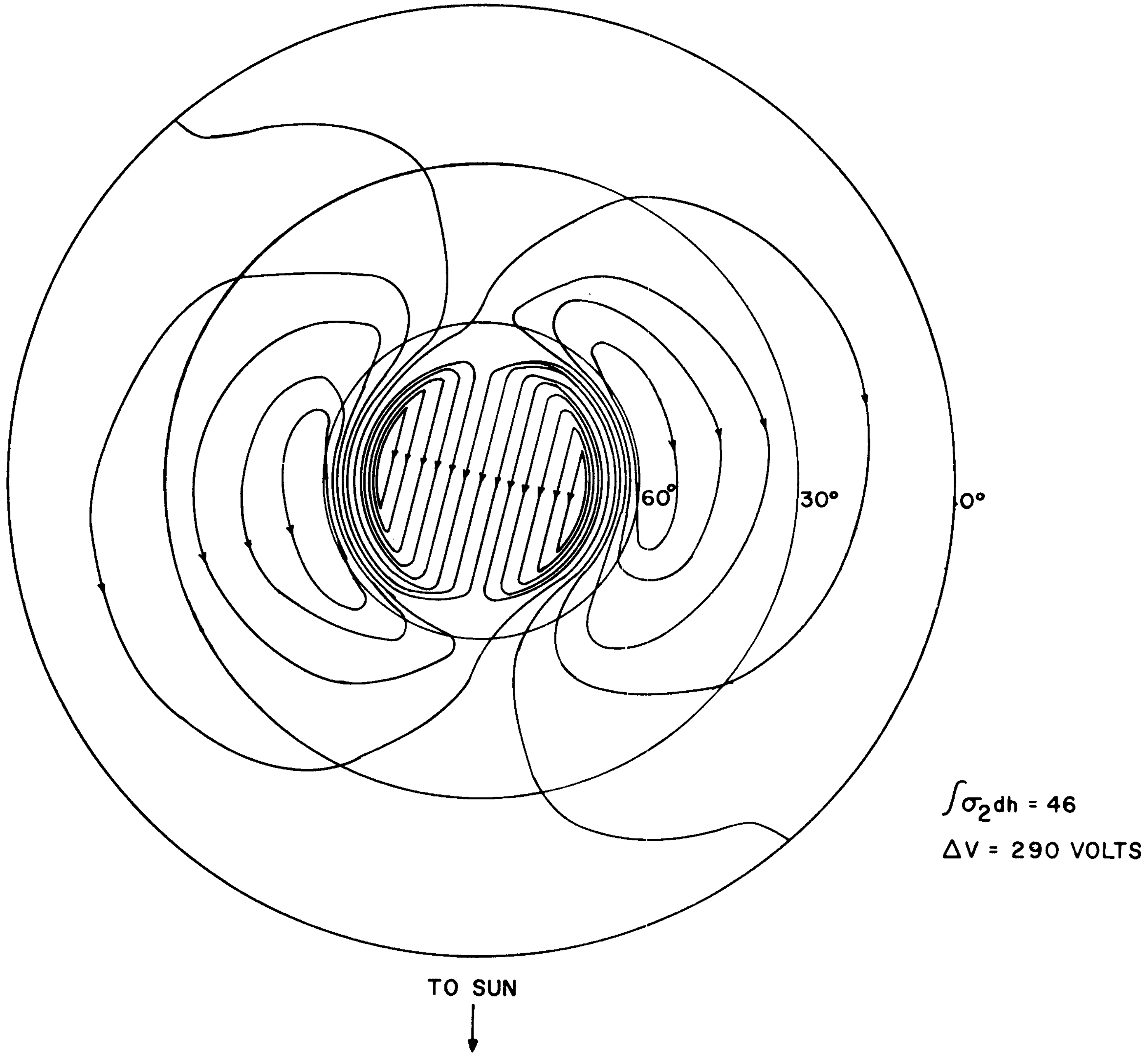
$\int \sigma_2 dh = 46$
CURRENT DENSITY (AMP / KM)
PHASE (HOURS)
LATITUDE (DEGREES)
S
E
E
S
70
60
50
40
30
20
10
12
6
24
18
12
90
80
70
60
50
40
30
20
10
0

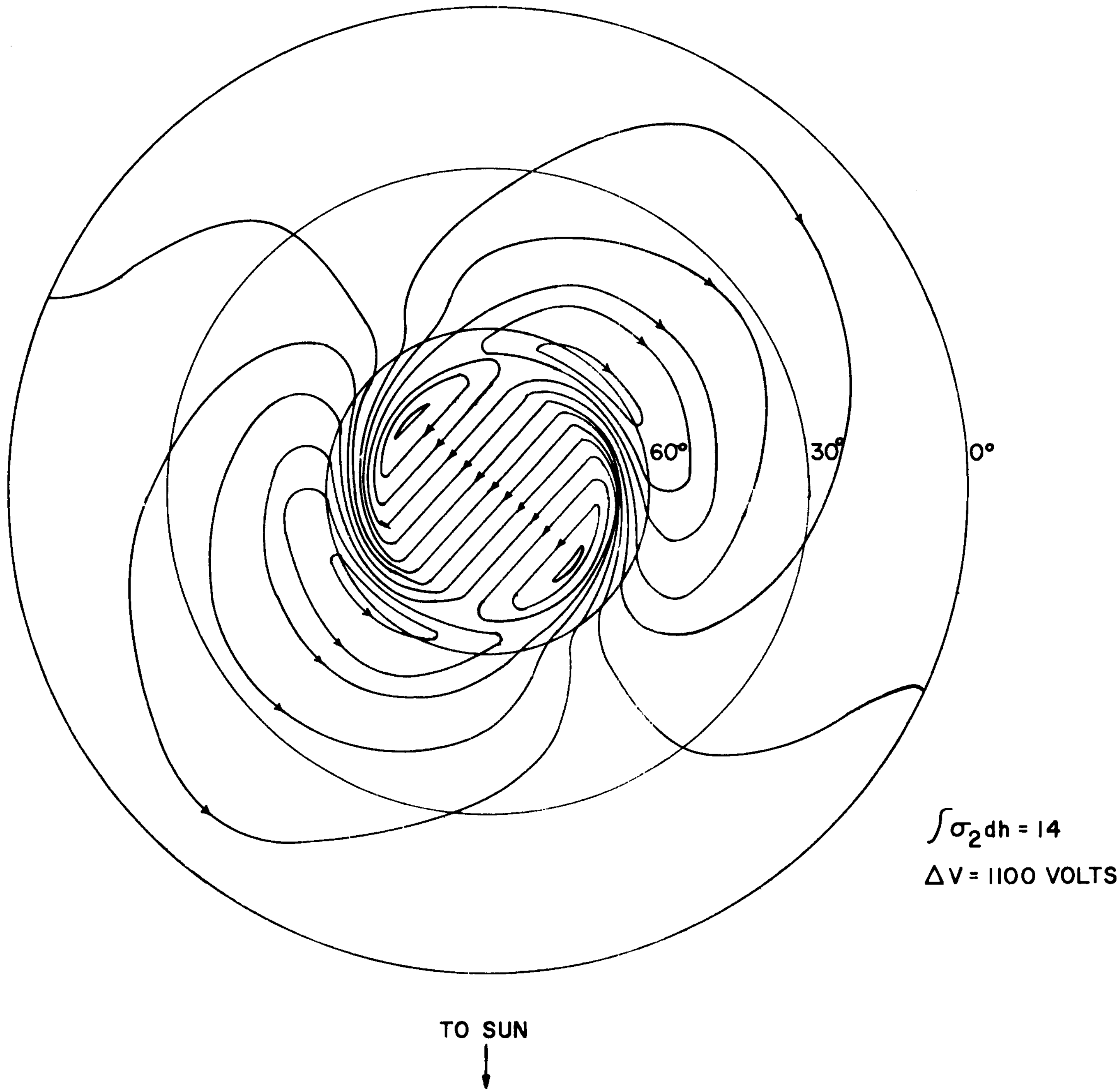


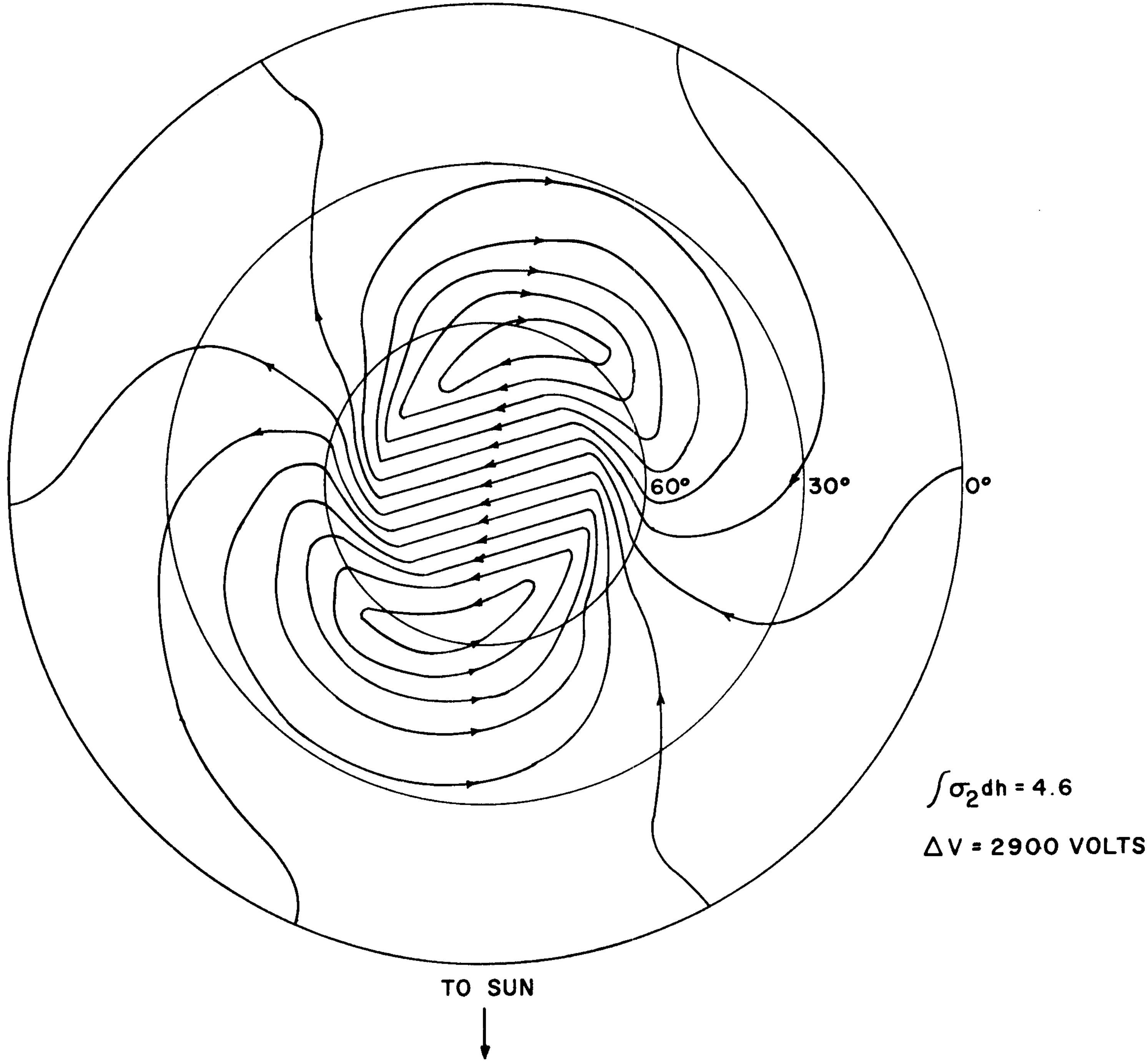
CURRENT DENSITY (AMP/KM)
PHASE (HOURS)
LATITUDE (DEGREES)
$\int \sigma_2 \, dh = 14$
S
E
E
S
50
40
30
20
10
12
6
24
18
12
90
80
70
60
50
40
30
20
10
0



$\int \sigma_2 dh = 4.6$
S
E
E
S
CURRENT DENSITY (AMP/KM)
25
20
15
10
5
PHASE (HOURS)
12
6
24
18
12
90
80
70
60
50
40
30
20
10
0
LATITUDE (DEGREES)


60°
30°
0°
$\int \sigma_2 dh = 46$
ΔV = 290 VOLTS
TO SUN


60°
30°
0°
$\int \sigma_2 dh = 14$
$\Delta V = 1100$ VOLTS
TO SUN


60°
30°
0°
$\int \sigma_2 dh = 4.6$
$\Delta V = 2900$ VOLTS
TO SUN